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A.C. Demagnetization of Unstable Triassic Keuper Marls from S.W. England

K. M. Creer

(Received 1959 April 20)

Summary

In 1953, 329 specimen disks taken from 35 rock samples covering a stratigraphical thickness of 170ft of Keuper Marl (Triassic) were collected from Sidmouth, Devonshire, and the results were described by Creer (1955, 1957a) in which an analysis of the resultant directions of natural remanent magnetization indicated the presence of three separate components, there called "*primary*", "*secondary*" and "*temporary*". Re-measurement of 65 of these rock disks from 26 of the 35 rock samples after storage in field free space for 3 years indicates that the *temporary* component of magnetization has decayed away. In further experiments, the same 65 rock disks were demagnetized in alternating magnetic fields of peak value up to 1200 oersteds. The directions of magnetization which were originally strung out along the great circle between the primary and secondary directions progressively coalesce into two opposed groups of magnetization as the intensity of the demagnetizing field is increased. It appears that in many cases the secondary magnetization has been almost completely destroyed and in the remainder of cases considerably reduced in comparison with the primary. The method of a.c. demagnetization is described.

1. Introduction

In 1953, while at the Department of Geodesy and Geophysics, Cambridge University, the author made a detailed study of the natural remanent magnetization of samples of Keuper Marl taken over a stratigraphical thickness of 170ft from the cliff to the east of Sidmouth, Devonshire, in S.W. England. 329 specimen disks cut from 35 samples were measured (Creer 1955, 1957a). The mean directions of magnetization of these 35 samples lay nearly along a great circle through the directions of the present axial dipole field and the mean Triassic field. The latter was calculated from the results of the work of Clegg &

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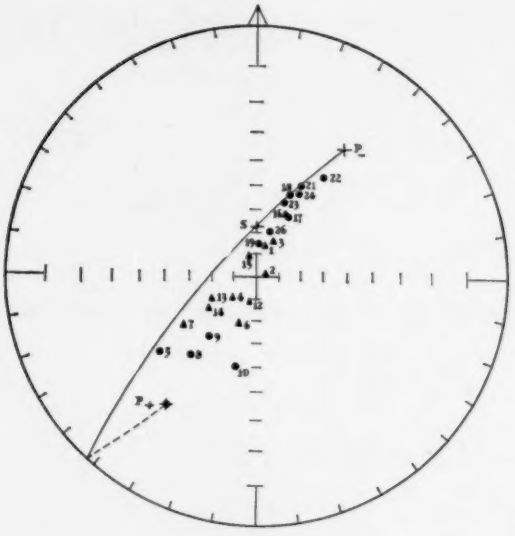

FIG. 1.—Original directions of magnetization of samples of Keuper Marl. Measurements made at Cambridge in 1953. Polar stereographic plot. Dots denote directions of north-seeking poles of n.r.m. of samples belonging to groups *N* and *R*, triangles those of samples belonging to groups *n* and *r* (see text). Sample numbers shown. Great circle $P_- SP_+$ connects primary and secondary directions of magnetization.

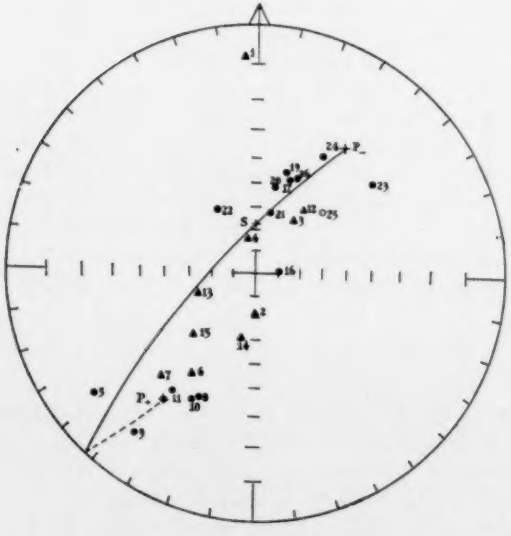

FIG.2.—Directions of magnetization of samples of Keuper Marl made in 1958 after storage in field free space for 2 years prior to measurement. Polar stereographic plot. Convention of plotting as for Figure 1.

others (1954) on Keuper sandstones and marls from 9 other sites in England where the natural remanent magnetization appeared stable.

The Keuper Marl would have been better named Keuper Mudstone for it is only slightly calcareous consisting of a fine silt with an admixture of clay minerals and small dolomite crystals. In many places it is structureless, closely resembling loess but at Sidmouth it is stratified at some levels probably due to deposition in temporary lakes.

When samples of this deposit are broken up a black ferromagnetic content can be separated. X-ray analysis shows this to be mostly haematite. The red cement may similarly be shown to be haematite.

2. Measurement after storage in field free space

The original measurements of directions of natural remanent magnetization made at Cambridge in 1953 are shown in Figure 1, the great circle being denoted by P_+ SP_-. For each sample the individual disk directions of magnetization have been combined in order to eliminate experimental errors and the points in Figure 1 represent mean directions for each of the 26 samples. (The term "sample" is used for each piece of rock taken from a field exposure. From each sample a number of cylindrical disks of diameter 3·75 cm and thickness 0·8 cm are cut; and measurements are made and experiments carried out on these.)

However, the inclinations of the natural remanent magnetizations are rather too steep; the points lie closer to the centre of the stereogram than the great circle. The probable explanation of this is that the disks were cut in all cases with their axes perpendicular to the bedding and were stored in the Earth's magnetic field before measurement with their bottom faces downwards but in random azimuths. Thus the vertical component of temporary magnetization was not meaned out in the original measurements, although the horizontal components were. This downward vertical component explained the effect observed in the stereogram in Figure 1.

The results obtained on remeasurement of these disks at Newcastle in 1957 after storage in zero field for about 3 years show that this explanation is satisfactory. As seen in Figure 2 the mean sample directions are now more evenly distributed on either side of the great circle P_+ SP_-.

3. Procedure of a.c. demagnetization of rocks

A.c. demagnetization has been successfully applied by Brynjolfsson (1957), Creer (1958) and As & Zijderveld (1958) to remove unstable components of magnetization of igneous rocks without appreciably affecting the primary components. To achieve this demagnetization, it has been found necessary to use a.c. fields of peak value about 100–200 oersteds which is rather less than the coercivities of the igneous rocks (see Nagata 1953). Now a back field of the same order as the coercivity is required to remove the saturated remanances and for sediments Creer (1955, 1957b) has found this to be about ten times greater varying from 2 000 to 10 000 oersteds and to be 2 700 oersteds for the Keuper Marls. Hence it was thought that demagnetizing fields up to about 2 700 oersteds would not completely destroy the primary (fossil) component, while fields of about 1 000 oersteds peak value might be necessary to demagnetize the secondary component.

The coil system used in these experiments and in those of Creer (1958) was built and designed by Parry (1957) and consists of 2 000 turns of 17 swg wire wound in 22 layers, and has an internal diameter of 13·6 cm and an external diameter of 20·4 cm and is 13·6 cm long. It has a self inductance of 0·4 henries and is tuned to the 50 c/s mains supply by connecting in series with a bank of condensers, the exact capacitance being determined by trial and error during tuning tests. The coil was calibrated by measuring the output of a small search coil placed at its centre and was found to produce 200 oersteds (peak value) per ampere (r.m.s.). The 240 V mains will give a maximum current of 10 A (r.m.s.) and hence the coil is capable of producing an alternating magnetic field of 2 000 oersteds peak value.

Clegg (private communication) and As & Zijderveld (1958) report random magnetization of some specimens after a.c. demagnetization and claim that this is due to the presence of unsymmetrical even harmonics in the mains supply, producing a magnetization in the specimens so treated. Creer (1958) demagnetized specimens about three mutually perpendicular axes in both directions and this effect was not marked. However in strong alternating magnetic fields, results were not always entirely satisfactory so a device was built in which 4 sedimentary rock disks 3·5 cm diameter and 0·8 cm thick or 1 igneous core 3·5 cm diameter and 3·5 cm thick could be rotated simultaneously about two perpendicular axes while in the demagnetizing field. These axes are both perpendicular to the axis of the coil and have speeds of rotation in the ratio 1 : 1·1 so that any direction in the sample performs a Lissajous figure during rotation and all directions in the rock are demagnetized. The speed of rotation can be varied between about 1 and 3 revolutions per second, which is roughly 50 to 20 times slower than the frequency of the main current supply. This device thus prevents the specimens acquiring an i.r.m. due to d.c. components of the origin mentioned above. Care was taken to ensure that the speeds of rotation of the specimen were not sub-harmonics of 50 c/s.

The normal practice in a.c. demagnetization is slowly to decrease the demagnetizing field to zero from the chosen maximum value. It is important that the rate at which this field is decreased is even and slow compared with the speed of rotation of the specimen. To ensure a smooth and slow fall off in demagnetizing field an electrolytic resistance has been introduced in the circuit. It consists of a tube 10 cm in diameter and about 1 m long containing copper sulphate solution. One copper electrode is fixed to the bottom and another to a probe which is gradually pulled upwards through the electrolyte by means of a string and pulley system powered by an electric motor which turns at 30 revolutions per hour. It is arranged, by gradually increasing the speed of withdrawal of the probe as the electrode separation is increased, that the resistance of the electrolyte increases at a uniform rate and hence the a.c. current through the coil is reduced steadily and uniformly. It requires 10 min to reduce the demagnetizing field from about 1 000 oersteds to zero, and the value of the current passing through the demagnetizing coil is registered on a pen recorder. A photograph of the apparatus is shown in Figure 3.

In earlier experiments in which the demagnetizing field was gradually decreased to zero by slowly pulling the coil away from the specimen along non-magnetic roller bearings, it proved impossible to withdraw the coil sufficiently slowly and smoothly because of uneven friction. (The field intensity drops by a factor of about $\frac{1}{3}$ during the initial movement of 6·8 cm, the half length of the coil.) Hence this procedure was abandoned in favour of the variable electrolytic resistance, but the

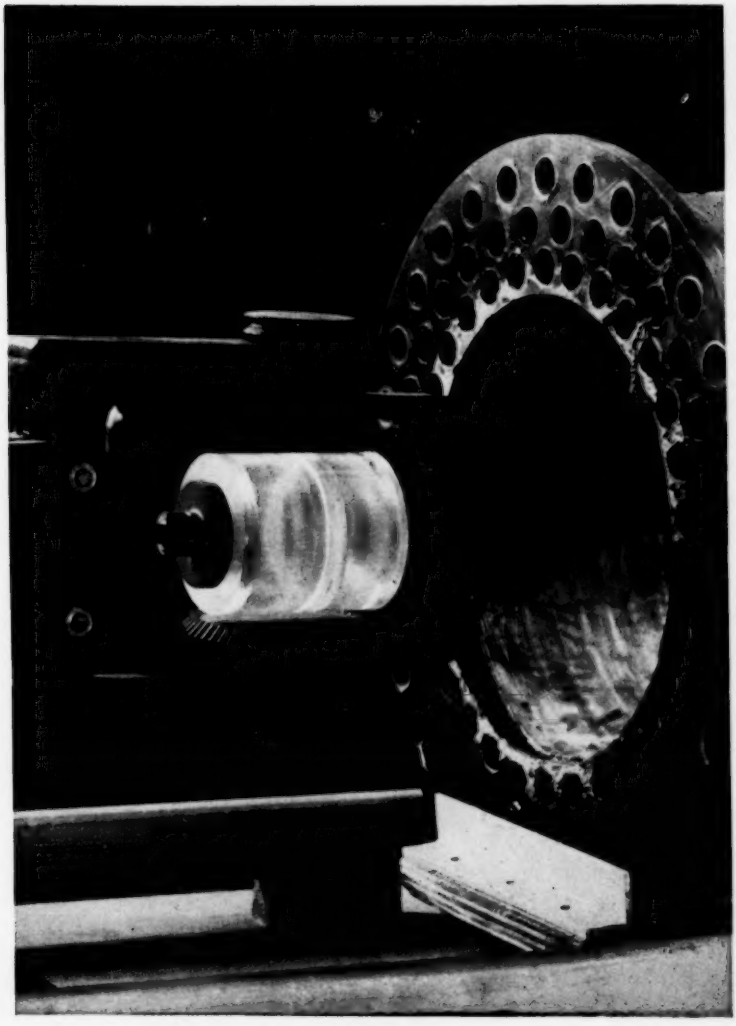

FIG. 3.—Photograph of demagnetizing coil and specimen holder.

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latter is not satisfactory when used for more than a few minutes at a time when passing currents of greater than 6 A and this limited the intensity of a.c. field applied in these experiments to 1 200 oersteds. It has, however, since been found possible to reduce the field sufficiently smoothly by suspending a cradle 1 m long containing the coil from the ceiling by six threads such that it can move only in a direction parallel to its axis and slowly pulling it away automatically.

4. Demagnetization experiments on Keuper Marls

4.1. Experimental procedure

The 65 rock disks were demagnetized by alternating magnetic fields of peak value 400, 600, 800, 1 000 and finally 1 200 oersteds. The mean direction of magnetization of each of the 26 rock samples was found by averaging the directions of magnetization of its 2 or 3 component disks, the effect of stray magnetizations acquired during "cleaning" being reduced considerably in this way. Stereographic projections of the results are shown in Figures 4 to 9.

4.2. Classification of results

An inspection of Figure 2 and Figures 4 to 9 shows that it is possible to classify the rock samples into four groups: (i) *group N* consisting of samples the directions of magnetization of which are centred about the mean direction of magnetization P_- of stable normally magnetized Triassic rocks, (ii) *group R* consisting of samples the directions of magnetization of which progressively coalesce around the mean direction of magnetization P_+ of stable reversely magnetized Triassic rocks as the value of the demagnetizing field is increased and (iii) *groups n and r* consisting of samples which are apparently not completely cleaned by the treatment described. Included here also are two samples the directions of magnetization of which do not change in an obviously systematic manner.

In Figures 2, 4, 5, 6, 7, 8 and 9 points representing directions of magnetization of groups *N* and *R* are shown by dots or open circles and those of groups *n* and *r* by full or open triangles.

Paths traced by successive directions of magnetization of samples at 0, 400, 600, 800, 1 000 and 1 200 oersteds respectively during the experiments are shown in Figure 13. Theoretically, if the directions are resultants of primary and secondary components these paths should follow great circles. That they do not is due to stray magnetizations picked up during the cleaning process rather than to errors in measurement. One marked example of this is given by sample 12 (group *r*) at 1 000 oersteds. Sample 26 (group *N*) shows no significant change in direction, though group *N* as a whole shows a significant change in average direction of magnetization during the experiment as is shown in Figure 11. Sample 2 (possibly group *n*) shows no significant change, though sample 10 (group *R*) gives a path approximating to the great circle P_-SP_+.

4.3. Discussion of results

4.3.1. *Stratigraphical succession of samples.*—The rock samples have been numbered in order of stratigraphical succession as indicated in Table 1, from which the measurements described in this paper may be correlated with the original measurements published by Creer (1957a).

FIG. 4.—Directions of magnetization of samples of Keuper Marl after demagnetization in a.c. magnetic field of peak value 400 oersteds. Dots and full triangles represent directions of north-seeking poles of remanent magnetization with downward inclinations, circles and hollow triangles, directions of north-seeking poles of remanent magnetization with upward inclinations.

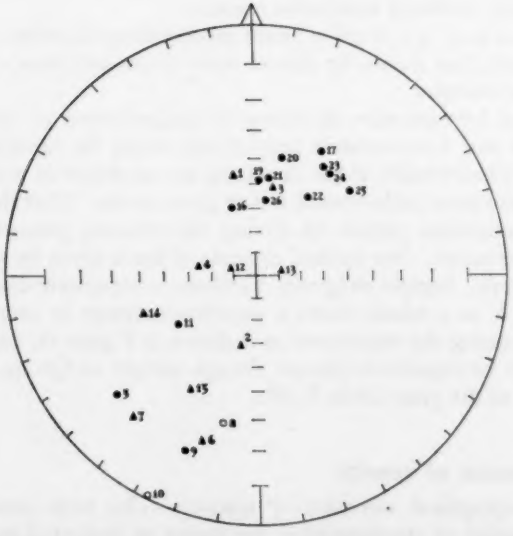

FIG. 5.—Directions of magnetization of Keuper Marl samples after demagnetization in a.c. magnetic fields of peak value 600 oersteds. Plotting convention as for Figure 4.

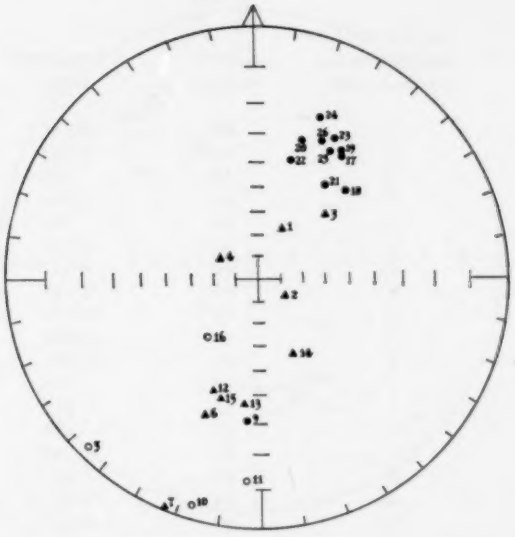

FIG. 6.—Directions of magnetization of Keuper Marl samples after demagnetization in a.c. magnetic fields of peak value 800 oersteds. Plotting convention as for Figure 4.

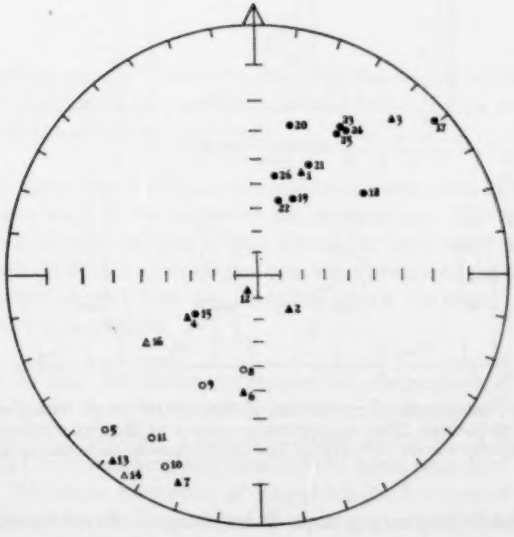

FIG. 7.—Directions of magnetization of Keuper Marl samples after demagnetization in a.c. magnetic fields of peak value 1 000 oersteds. Plotting convention as for Figure 4.

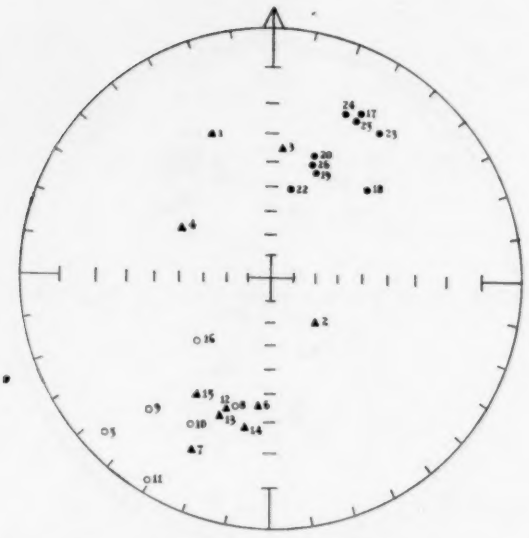

FIG. 8.—Directions of magnetization of Keuper Marl samples after demagnetization in a.c. magnetic fields of peak value 1 200 oersteds. Plotting convention as for Figure 4.

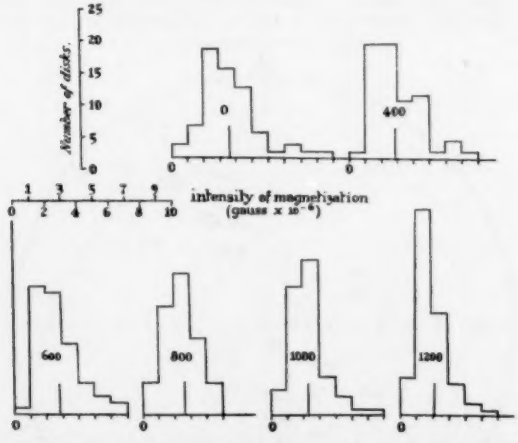

FIG. 9.—Histograms of intensities of magnetization of rock disks from samples of Keuper Marl at successive stages of the experiment. Mean intensity shown by vertical line drawn inside each histogram.

Those samples falling into groups *R* and *r* are confined to the middle 36·5 ft of the cliff, those falling in group *N* to the upper 36 ft. This group may continue stratigraphically upwards beyond the topmost horizon sampled because a sample from the 170 ft horizon previously studied (Creer 1957a) had a virgin direction of magnetization N 28 E + 40°. Samples falling into group *n* come from the bottom 35 ft of the cliff.

Table 1

Stratigraphic level	Sample number	Direction of magnetization after cleaning
145 ft	26	*N*
143·5	25	*N*
143	24	*N*
142·5	23	*N*
219	22	*N*
127·5	21	*N*
115·5	20	*N*
111·5	19	*N*
111·0	18	*N*
109·5	17	*N*
108·5	16	*R*
106	15	*r*
105	14	*r*
103	13	*r*
101	12	*r*
91·5	11	*R*
91·0	10	*R*
90·5	9	*R*
90	8	*R*
89·5	7	*r*
88·5	6	*r*
77	5	*R*
72	4	?
35	3	*n*
15	2	?
0	1	*n*

Samples comprising groups *N* and *R* come from the well stratified beds which appear to be water-deposited in contrast to those from groups *n* and *r* which come from the more massive loess type beds.

4.3.2. *Intensity histograms.**—Figure 9 consists of histograms of the intensities of individual disks at each of the stages of the experiment. The mean intensity of magnetization for all the samples is also shown for each stage and it may be observed first that it decreases monotonically as the strength of the demagnetizing field is increased, and second that the deviation about the mean direction also decreases as the field is increased.

4.3.3. *Groups N and R.*—Mean intensities of magnetization at successive stages of the experiment have been calculated for samples 17–26 inclusive comprising group *N* and for samples 5 and 8–11 from group *R*. These, with their standard deviations σ and standard deviations of the mean (standard errors) σ_m are listed in Table 2. The mean intensities of magnetization for each of groups *N* and *R* are plotted as functions of the peak value of demagnetizing field in Figure 10 from which it is seen that the intensities of magnetization of groups *N* and *R* are not significantly different from one another after cleaning although the intensity of magnetization of group *N* was considerably greater than that of group *R*, after

* Due to an error, all intensities of magnetization given in Creer (1957a) for the original magnetization are too low by a factor of 10.

Table 2

Mean intensities of magnetization of Keuper Marl samples on demagnetization

units: gauss $\times 10^{-6}$

Peak value of demagnetizing field (oersteds)	Normal (Group *N*)			Reversed (Group *R*)			Others (Groups *n* and *r*)		
	M	σ	σ_m	*M*	σ	σ_m	*M*	σ	σ_m
0	4·37	1·15	0·24	2·98	1·11	0·37	3·45	2·38	0·47
400	3·77	1·33	0·27	2·37	0·75	0·23	2·51	1·35	0·27
600	3·02	1·23	0·26	2·63	0·76	0·23	2·65	1·46	0·29
800	2·89	0·92	0·19	2·43	0·54	0·16	1·82	0·94	0·20
1 000	2·54	1·18	0·25	2·39	0·77	0·22	2·04	1·18	0·26
1 200	2·22	1·00	0·21	2·39	0·88	0·25	1·77	0·69	0·14

storage in field free space and before cleaning. Mean directions of magnetization of samples from groups *N* and *R* have been calculated after each demagnetization. These are shown in Figure 11 where S, the assumed direction of the secondary

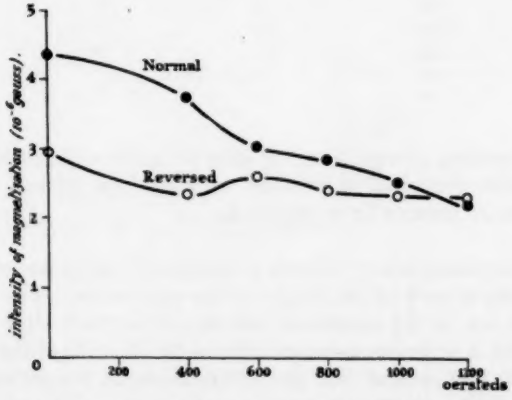

FIG. 10.—Average intensities of magnetization of all normally magnetized samples (17 to 26, i.e. group *N*) and of reversely magnetized samples (5, 8, 9, 10 and 11, from group *R*) at successive stages of the experiment.

component, is that of a geocentric axial dipole field at Sidmouth and P_- and P_+ the assumed normal and reversed directions of the primary component are the poles of the mean axis of magnetization of stable Keuper sandstones studied by Clegg & others (1954). Successive mean directions of magnetization of both groups move along paths close to the great circle P_- SP_+ indicating that the secondary component is demagnetized more readily than the primary. The angles between the resultant directions of magnetization and the assumed primary and secondary directions were measured graphically and the intensities of the

primary and secondary components were then calculated for each stage. In Figure 12 these are shown as functions of the peak value of the demagnetizing field. From these graphs the secondary component is seen to be much "softer" than the primary, the former being almost completely destroyed by a.c. fields of 800–1 000 oersteds and the latter retaining about three-quarters of its initial value after having been subjected to demagnetization in an a.c. field of 1 200 oersteds. Assuming that the angles between the resultant directions of magnetization and P_ and S respectively can be estimated graphically to 2°, the estimated error for the relative intensities of components *P* and *S* works out at about 15 per cent. This probably explains the scatter of the *P* and *S* intensities shown in Figures 12a and 12b.

The successive changes in direction of magnetization of some individual samples from groups *N* and *R* are shown in Figure 13.

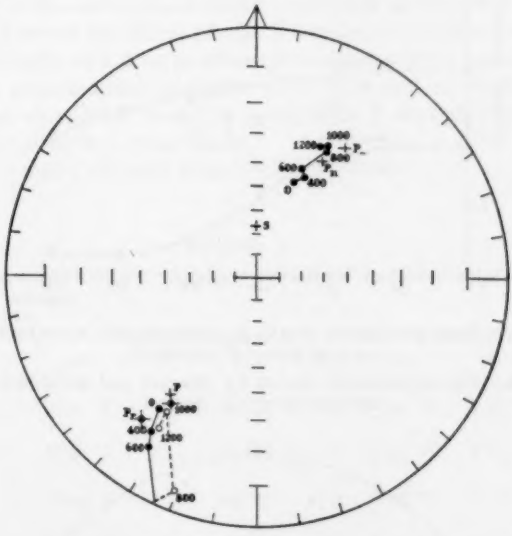

FIG. 11.—Paths followed by average directions of magnetization of groups *N* and *R* during the experiment. P_-, P_+ normal and reversed directions of Triassic axis of magnetization. P_n and P_r mean directions of magnetization of normally and reversely magnetized Keuper Sandstones studied by Clegg & others (1954). S direction of dipole field along present geographic axis at Sidmouth.

4.3.4. *Group r.*—The samples of this group come from the middle of the cliff. The directions of magnetization after storage in zero field (Figure 2) are all steeply inclined to the bedding planes, close to the great circle P_-SP_+ and near S, the direction of the secondary component. This suggests that the ratios between the intensities of the secondary and primary components are greater than in groups *N* and *R*. As the cleaning became progressively more severe the directions of magnetization of these samples moved along paths close to the great circle P_-SP_+ towards P_+ (for examples see Figure 13), but showed no sign of reaching the point P_+. Further cleaning in demagnetizing fields higher than 1 200 oersteds might align the resultant directions of magnetization of these samples more closely parallel to P_+ as in group *R*. However, the intensity of magnetization is generally

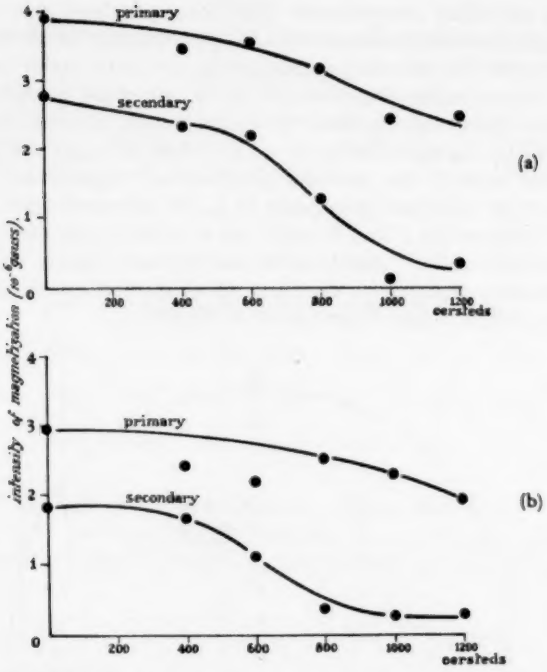

FIG. 12a.—Demagnetization curves for primary and secondary components of group *R* (reversed).

FIG. 12b.—Demagnetization curves for primary and secondary components of group *N* (normal).

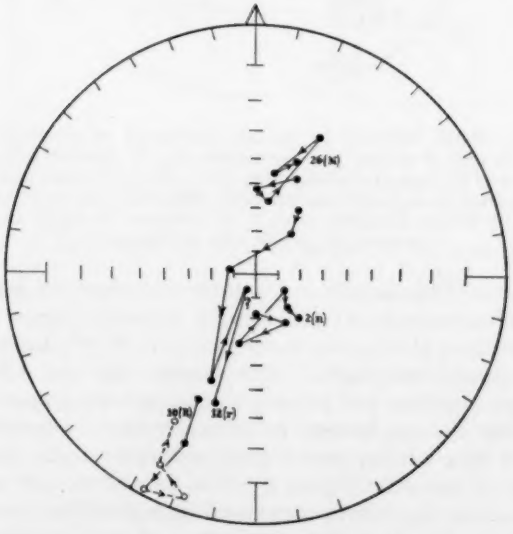

FIG. 13.—Successive directions of magnetization of representative samples (numbers and groups shown) after treatment in peak demagnetizing fields of 0, 400, 600, 800, 1 000 and 1 200 oersteds respectively.

lower than in group *R* and repeated readings of intensity and direction for disks subjected to the same treatment more than once are not so good as for group *R*. It is possible that the directions of magnetization after cleaning in fields of 1 000 or 1 200 oersteds represent the direction of the geomagnetic field when these beds become magnetized, but this is improbable because of the poor repeatability and because the *rR* zone is bounded at its upper and lower limits by *R* type samples (see Table 1). That samples of group *r* come from the loess type beds is the most probable reason for their different experimental behaviour from samples from group *R* which come from thinner water laid beds.

4.3.5. *Group n.*—Samples 1 and 3 show similar characteristics to those of group *N*, their directions of magnetization tending to move towards P_ as the demagnetizing field is increased. However, repeated measurements following further treatment in the same demagnetizing field lack repeatability as in group *r*. Samples 2 and 4 show no significant change in direction of magnetization on cleaning, although both suffered a drop in intensity to about half the initial value. From their response to experimental treatment alone it is difficult to decide whether these two samples should be placed in group *n* or *r*, though from their stratigraphical position group *n* is more likely.

Samples 1, 2, 3 and 4 all come from loess type beds.

Table 3

Directions of magnetization, standard deviations θ_0 and standard errors θ_m

Demagnetizing field (peak value in oersteds)	Normal (Group *N*)				Reversed (Group *R*)			
	D	*I*	θ_0	θ_m	*D*	*I*	θ_0	θ_m
0	N 22° E	+46°	14°·7	5°·2	S 36° W	+22°	15°·3	8°·8
400	N 26° E	+43°	16°·8	5°·6	S 34° W	+15°	19°·6	9°·8
600	N 25° E	+40°	14°·5	5°·1	S 32° W	+11°	32°·8	16°·4
800	N 29° E	+30°	9°·2	3°·1	S 18° W	+ 3°	26°·2	15°·2
1 000	N 29° E	+29°	19°·9	6°·6	S 34° W	−23°	20°·7	10°·3
1 200	N 27° E	+30°	14°·8	5°·2	S 33° W	−17°	16°·9	8°·4
Mean value for stable Triassic rocks	N 29° E	+34°	—	—	S 39° W	−16°	—	—

5. Conclusions

Demagnetization experiments on Keuper Marl samples have confirmed that the natural magnetization of these rocks consists of a *primary* component in the same direction as the stable n.r.m. of normally or reversely magnetized Keuper Sandstone together with a magnetically softer *secondary* component. The latter requires an alternating field of peak value greater than 800 oersteds to destroy it, and is therefore considerably harder than the soft component of unstable igneous rocks which is completely demagnetized by a.c. fields of peak value 100–200

oersteds. It is also considerably harder than the stable magnetization found in igneous rocks which is reduced to roughly half by a.c. fields of peak value 300–400 oersteds (Creer 1958).

The primary component of this sediment is extremely hard and retains about three-quarters of its natural value after demagnetization in a.c. fields of peak value 1 200 oersteds.

The 145 ft of Keuper Marl studied appear to consist of an upper normally magnetized zone more than 35 ft and possibly more than 60 ft thick, a central reversely magnetized zone of thickness 36·5 ft and possibly a lower normally magnetized zone. No chemical or mineralogical differences between the normally and reversely magnetized samples have been detected, both types of zone containing water-laid and loess type deposits.

If the reversed magnetization is due to a reversal of the main geomagnetic field it should be possible to infer something of the way the reversal takes place and of the time required for a reversal to occur. However, although the estimated stratigraphical separation between the top sample (No. 16) of the reversed zone and the bottom sample (No. 17) of the upper normal zone is only 1 ft these samples come from different beds. Further, a horizontal distance of 10 ft separated samples 16 and 17 and this was covered by scree. A fault between these two points appeared unlikely but the possibility of a small fault of a few feet throw must be borne in mind and therefore the time taken for the reversal to occur cannot be estimated. Also there is no way of estimating the time interval between the deposition of the beds.

The transition between the lower normal zone and the base of the reversed zone does not appear to be nearly so sharp, though sampling at this horizon was not carried out so intensively and the loess type of rock encountered here is not very suitable for palaeomagnetic work because of the possible existence of a comparatively hard secondary component.

The thickness of a complete reversed zone of about 36 ft is probably accurate only to 5 ft as the beds could not always be followed along the base of the cliff due to rock falls. However, no fault of more than a few feet was encountered. Little is known of deposition rates of sediments and no geologist acquainted with this formation is prepared to venture a reliable estimate of the time required for the deposition of this 35 ft. Further, since the origin of magnetization was probably chemical, rates of deposition may not have any direct bearing on the time required for magnetization to take place.

Further field work is planned in this region to study in more detail the horizons in which the reversals occur. When the field work was originally done in 1953 the intention was merely to measure the mean direction of n.r.m. of this formation and for the more detailed studies now planned it will be necessary to make much more thorough collections. More detailed studies on the exposures of stable Keuper sandstones visited by Clegg & others (1954) are also planned by the present author as in their original work only one or two rock samples were collected from each locality. A more intense sampling over a considerable stratigraphical thickness might possibly reveal a similar pattern of reversals to those found at Sidmouth.

If this occurs and if the reversals are not of equal duration it may be possible to correlate the different isolated outcrops of Keuper rocks in England, which is not possible by normal geological methods.

6. Acknowledgments

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Anomaly of Geomagnetic Variations in Japan

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Summary

The anomalous features found in geomagnetic variations of various kinds, rapid and slow, are described on the basis of observations at well-distributed observatories in Japan. Since the anomalies are proved to be caused by the induced magnetic field arising from inside the Earth, the underground conditions responsible for the anomalies have been intensively examined. At the moment, the best way to account for the anomaly would be to assume that a conducting circuit, which comes up from the deep, forms a loop, about 1 000 km in length and 200 km in width, beneath Japan. It is also supposed that a wedge-shape penetration of the upper layer of the mantle must exist there. The depth of the bottom of the penetrating part is estimated at 700 km. These complicated structures seem to have an important bearing on the geophysical interpretation of the relationship between occurrence of deep-focus earthquakes, orogenic activity, trend of volcanic zones and so on.

1. Introduction

Japan is one of the countries which are covered by a network of well-distributed magnetic observatories. Through the observations that have been carried out at these observatories, a very interesting, though rather anomalous, feature of geomagnetic variations has recently been brought out. Since it seems likely that the anomalous behaviour has an important bearing on the physical conditions underneath Japan, it is intended in this paper to report synthetically the observational results, together with their possible interpretation, though a number of papers has already been published as the investigation progressed (Rikitake, Yokoyama & Hishiyama 1952, 1953a, 1953b, 1953c, 1953d, Rikitake & Yokoyama 1953, 1954, 1955, Rikitake, Yokoyama & Sato 1956, Rikitake, Yokoyama, Uyeda, Yukutake & Nakagawa 1958, Rikitake, Uyeda, Yukutake, Tanaoka & Nakagawa 1959).

What we call here the anomaly of geomagnetic variations differs from the anomaly of the geomagnetic field itself. We observe no marked anomaly of the geomagnetic field and its secular variation in Japan. Whenever we observe geomagnetic disturbances such as bays or magnetic storms, however, a local anomalous distribution of the disturbance field is observed in Japan. Even in the case of daily variation (Sq), an unusual tendency of distribution has been found.

It has long been noticed that, at times of geomagnetic bays and polar magnetic storms, changes in the vertical component of the Earth's magnetic field (ΔZ) are

unusually large at Kakioka Magnetic Observatory, amounting to more than 50 per cent of changes in the horizontal component (ΔH). No intensive investigations to clarify such an anomalous phenomenon had been tried, however, until Rikitake and his collaborators set up a number of temporary magnetic observatories around Kakioka area. They confirmed that the large amplitude of ΔZ is only observed in a certain area spread over the central part of Japan. In the anomalous area in which $\Delta Z/\Delta H$ exceeds 0·5, marked parallelism is observed between ΔH and ΔZ. According to the recent observations carried out on the islands in the south sea, the area seems likely to spread in a roughly elliptical shape, about 1000 km in length and 200 km in width, along the south coast of the main island of Japan. In the central part of the area, $\Delta Z/\Delta H$ slightly exceeds 1.

Irregular distributions of ΔZ at times of sudden commencements of magnetic storms (*s.s.c.*) have been sometimes reported. Some of such irregularities are ascribed to electric currents induced in the sea. The reversal of sign of Z at British and French magnetic observatories has been well known. Barber (1948) has explained the reversal as the influence of electric currents induced in the English Channel. From various reasons, however, it seems hardly possible to account for the Japanese anomaly as the effect of electric currents induced in the sea.

A similar sort of anomaly associated with geomagnetic bays and polar magnetic storms has been also found in Germany. A reversal of sign of ΔZ is found at observatories in North and South Germany. According to German colleagues (Meyer 1951, Bartels 1954, Kertz 1954, Fleischer 1954, Wiese 1954, 1956, Siebert & Kertz 1957), the anomalous distribution of disturbance field is likely to be caused by induced electric currents flowing in the east–west direction along a nearly straight line underneath North Germany, the currents being likely to flow at depths of 100–200 km.

Since the above-mentioned findings in Japan and Germany seem to be of geophysical importance, further investigations have been intensively conducted not only on geomagnetic bays and polar magnetic storms but also on other sorts of geomagnetic variations, rapid and slow. It is surprising that we have found anomalies of *Sq* as well as main phase of magnetic storm (*Dst*) of utterly different character in Japan.

Close examinations of the *Sq* make it clear that the time of the beginning of morning decrease as well as that of maximum decrease of the vertical component of the *Sq* occurs one or two hours earlier at observatories situated in Central Japan than at any other observatories in and around Japan. From an analysis of *Dst*, it is also found that ΔZ in Japan is considerably larger than its mean value usually observed in the middle latitude. It has been also proved that all these anomalies are caused by the magnetic field arising from inside the Earth. Since *Sq* and *Dst*, which change more slowly than geomagnetic bays or polar magnetic storms, induce electric currents in a deeper part of the Earth, there is a possibility that a synthetic view of the electrical properties beneath Japan from the top of the mantle down to a depth of several hundred kilometres can be formed by summing up all the analyses of geomagnetic variations.

The electrical state of the Earth's interior has been hitherto investigated by applying the theory of electromagnetic induction to the mean state of the Earth as a whole. However, it is possible to add something new to the existing knowledge of the Earth's interior by studying the local anomalous behaviour of geomagnetic variations as found in Japan. A hypothetical interpretation of the

relationship between the orogenic zone running along the Kamchatka–Japan–Mariana arc and the underground conditions as inferred from the present study is attempted in the last part of this paper.

2. Magnetic stations

Magnetic data for the present study have been supplied from many magnetic stations, permanent and temporary. Although they do not always cover the same observation period, their localities are summarized in Table 1.

Table 1

Magnetic stations

No.	Station	Geographic latitude deg. N	Geographic longitude deg. E
1	Memambetsu	43·9	144·2
2	Onagawa	38·4	141·5
3	Maze	37·7	138·8
4	Hosono	36·7	137·8
5	Komoro	36·3	138·4
6	Kakioka	36·2	140·2
7	Kanosan	35·3	140·0
8	Aburatsubo	35·2	139·6
9	Simosato	33·6	135·9
10	Hachijo	33·1	139·8
11	Aso	32·9	131·0
12	Aogasima	32·5	139·8
13	Kanoya	31·4	130·9
14	Kikaigasima	28·6	129·9

3. Anomaly of geomagnetic bays and polar magnetic storms

The local anomalous behaviour of geomagnetic variations in Japan can be best seen in the cases of geomagnetic bays and polar magnetic storms which usually last a few hours. In Figure 1 is illustrated the distribution of changes in the three geomagnetic components at the time of a bay that occurred on 1958 April 18. Looking at the curve of each component, it is noticeable that the ΔH and ΔD (change in declination) curves run regularly throughout all the stations. We observe, however, a marked tendency of ΔZ to be surprisingly large along the south coast of the main island in Central Japan. It is striking to find that ΔZ curves at stations No. 5 and No. 8, the distance between these two stations amounting to only 170 km, differ so much from one another.

The distribution of ΔZ at the maximum epoch of the bay is shown on the map in Figure 2. In making up the equal-variation lines of ΔZ, results of many previous observations at various stations are also taken into account though we made no observation there this time. Figure 2 makes it clear that the anomalously large ΔZ is only observed in a limited area of roughly elliptical shape. At observatories in the area a linear correlation between ΔH and ΔZ has been well established by the previous observations, the mean $\Delta Z/\Delta H$ amounting to, for instance, 0·56 and 0·62 at stations No. 6 and No. 8 respectively. The ratio sometimes exceeds 1 at station No. 9. Outside the anomalous area, the parallelism between ΔH and ΔZ does not hold.

In order to seek the cause of such a complicated disturbance field, the field should first be separated into the parts arising from outside and inside the Earth. From such an examination, if it is performed satisfactorily, we may be able to see whether or not the magnetic field originating from inside the Earth is responsible for the anomaly. With the aid of magnetic data kindly sent from many observatories all over the earth, the separation was actually made by Rikitake and others (Rikitake, Yokoyama & Hishiyama 1953c) for a marked polar magnetic storm on 1936 June 19, when we had a number of temporary stations in Japan, Manchuria

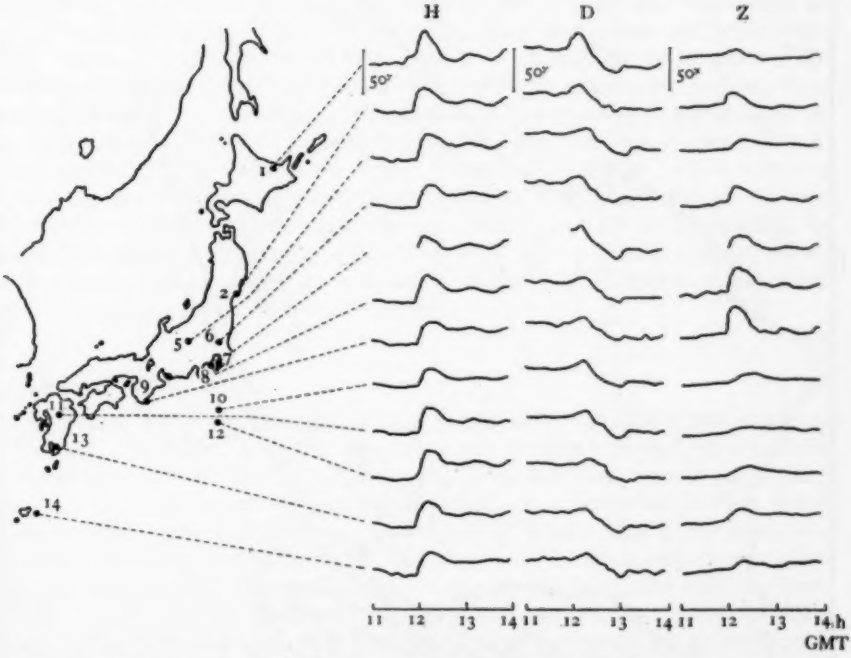

FIG. 1.—The changes in the three geomagnetic components at the twelve Japanese stations at the time of the geomagnetic bay on 1958 April 18.

and Sakhalin. The result of analysis shows clearly that the part of external origin is distributed quite regularly, while, on the contrary, the internal part is distributed in an anomalous way centring in Central Japan. The distribution of the latter is hardly understood from the standpoint of the theory of electromagnetic induction within a uniform Earth. In so far as we consider that the magnetic field of internal origin is produced by electric currents induced within the Earth by the change in the external magnetic field, the only way to account for such a strange distribution seems to presume some particular conditions of electrical state beneath Japan.

It is also shown that, at times of positive bays, the external part of ΔZ is directed upwards in Japan while its internal part points downwards. Since the amplitudes of both the parts are about the same at observatories in North Japan and along the coast of the Sea of Japan, ΔZ is usually very small in those districts. ΔZ of internal origin is, however, definitely much larger and positive in the

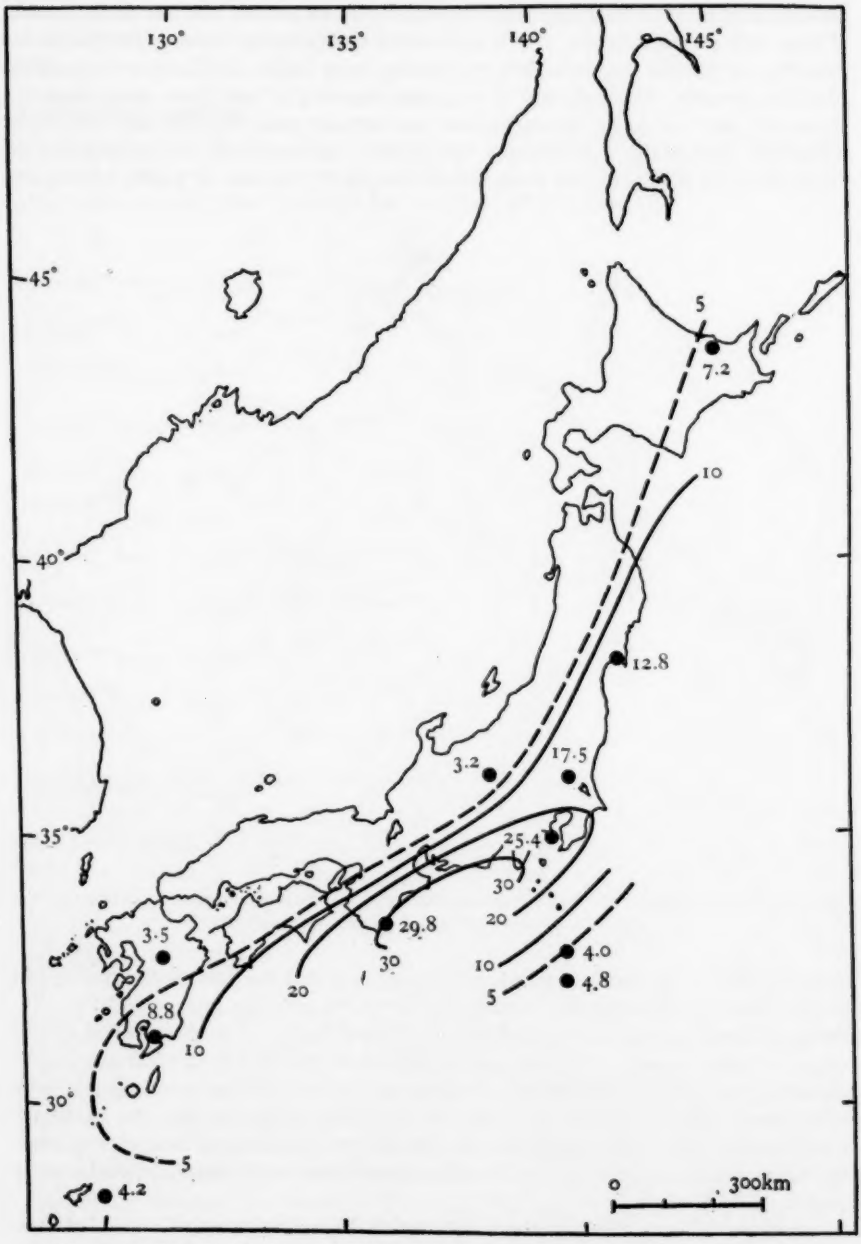

FIG. 2.—The distribution of ΔZ in units of gamma for the maximum epoch of the geomagnetic bay.

anomalous area, so that ΔZ of external origin is completely cancelled leaving some positive ΔZ there.

4. Anomaly of *s.s.c.*

As is briefly described in the last section, the anomalous behaviour of geomagnetic variations is best seen in Japan in the case of geomagnetic bays and polar magnetic storms. As for rapid variations such as *s.s.c.* and geomagnetic pulsation with duration-time or period of a few minutes, we also observe the anomalously large amplitude of ΔZ in Central Japan, though the distribution of ΔZ seems to be slightly different from that for geomagnetic bays and polar magnetic storms. It has been known that a rapid variation is apt to be affected by irregular distributions of the electrical conductivity near the Earth's surface. The question whether the particular distribution of ΔZ as observed in Japan might be due to the magnetic field produced by electric currents induced in the sea has been often raised. On the basis of the theory of electromagnetic induction within a conducting sheet (Rikitake & Yokoyama 1955), however, it has been shown that the effect of the sea would not be of the type which accounts for the anomaly. It is generally believed therefore that the anomaly of *s.s.c.* in Japan is closely correlated with the special underground condition though it is affected to some extent by the irregularities near the Earth's surface.

5. Anomaly of *Sq*

In Figure 3 is shown the *Sq* for each season observed at magnetic observatories Nos. 1, 6, 8 and 11. The curves are made by making averages of five of the quietest days in each calendar month during the period from 1952 to 1955. At a glance, we see that the vertical components of the *Sq* at observatories Nos. 6 and 8 show a particular phase difference in daytime variations, that is, the maximum decrease takes place about two hours earlier than at the other two observatories. Meanwhile, no marked phase difference is found in the curves for both the horizontal intensity and declination all through the observatories. Since the same tendency has been also noticeable in the analyses of *Sq* during the Second International Polar Year (Wilkins 1951, Rikitake, Yokoyama & Sato 1956), the particular phase-shift of the vertical component should be one of the characteristics of *Sq* in Central Japan. It is of interest that an anomaly of *Sq* of another character is found in the area in which we have been observing the anomalous behaviour of short-period geomagnetic variations.

With the aid of the geomagnetic data during the Second International Polar Year, the magnetic potential and vertical component of *Sq* arising from inside the Earth (W_i and Z_i respectively) are separated from those originating from outside the Earth (W_e and Z_e) at six observatories in the Far East by means of numerical integration method. At most observatories, the relationships between W_e and W_i and Z_e and Z_i are about the same as those expected for the average state of the Earth in which the distribution of the electrical conductivity has been well investigated so far (Chapman 1919, Chapman & Price 1930, Lahiri & Price 1939, Rikitake 1950a, 1950b, 1950c, 1951a, 1951b). It is found, however, that W_i and Z_i are definitely smaller than those for the averaged state of the Earth at Kakioka. Although only one station situated in Central Japan is available in the analysis, it is strongly suggested that the electrical state beneath Central Japan is different

from the average of the Earth as a whole. The anomaly of *Sq* thus deduced would be explained by assuming that the electrical conductivity probably at depths of 400 ~ 700 km is lower than the usual part of the Earth.

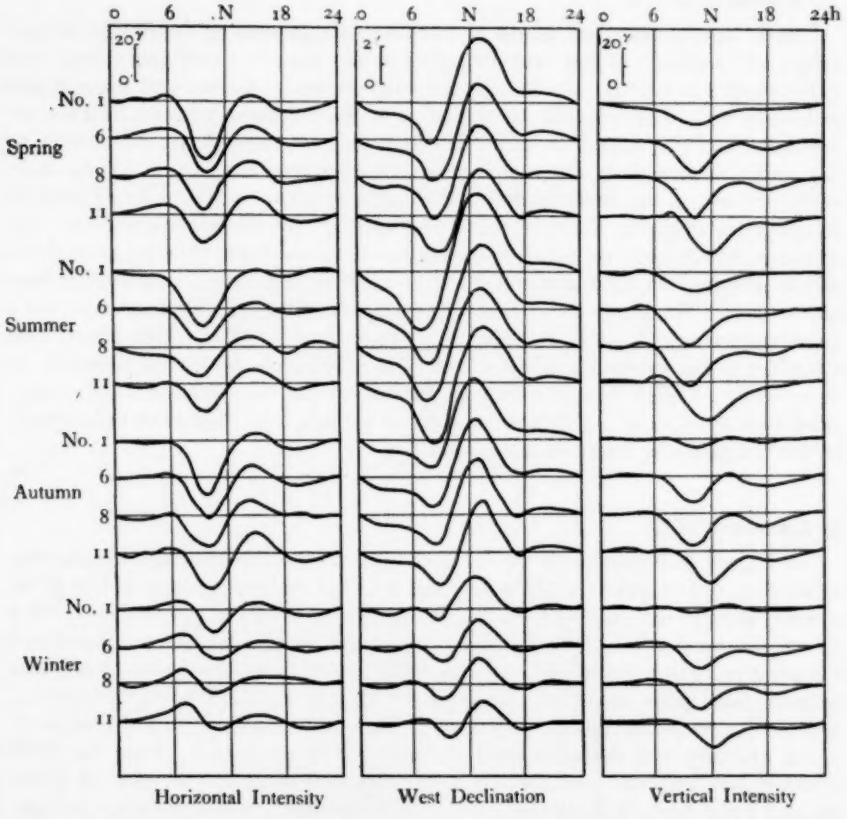

FIG. 3.—The *Sq* variation of the three geomagnetic components as obtained for respective seasons at the four Japanese observatories during the period 1952–55.

6. Anomaly of the main phase of a magnetic storm

The *Sq* observed in Japan seem to be affected very little by the underground conditions which cause the anomalously large ΔZ in the cases of geomagnetic bays and polar magnetic storms though it is anomalous in a different way. Since it is evident that the slower the magnetic change is, the deeper the electromagnetic induction effect penetrates into the Earth, the anomaly of *Sq* may reflect the conditions in the relatively deeper part of the Earth beneath Japan. It is hoped therefore to analyse the *Dst* part of magnetic storms, which varies more slowly than *Sq*, in order to make a further study of the deep interior of the Earth beneath Japan.

An analysis of the *Dst* of a magnetic storm that occurred on 1957 March 10 was made on the basis of the observation at nine observatories widely distributed

over Japan. The changes in the three geomagnetic components are shown in Figure 4. Short-period variations are eliminated from the curves by taking running averages of 13 hourly readings. Unlike geomagnetic bays or polar magnetic storms, no extraordinarily large amplitude of ΔZ can be seen in Central Japan.

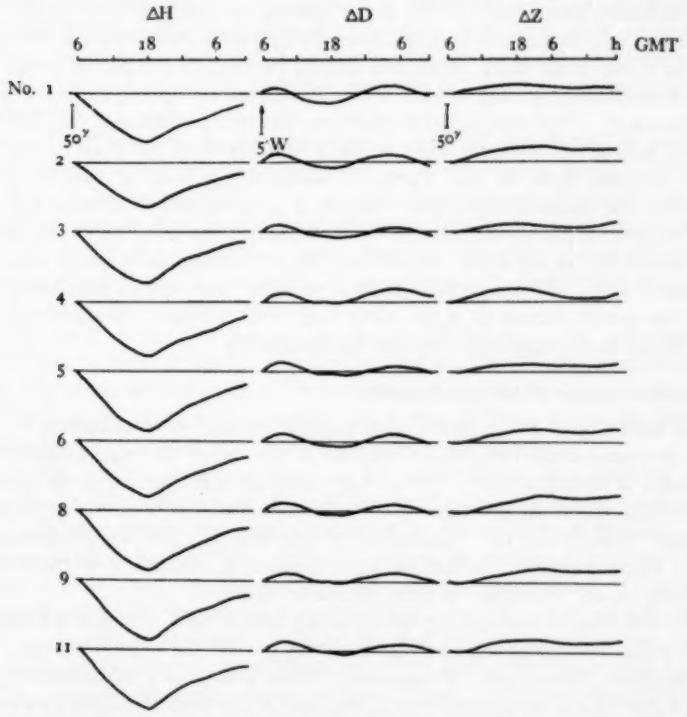

FIG. 4.—The *Dst* part of the magnetic storm on 1957 March 10. Short-period variations are eliminated.

Since the *Dst* seems to be distributed fairly regularly all through Japan, its magnetic potential may be expressed by

$$\Delta W = a\{(r/a)e + (r/a)^{-2}i\}P_1(\cos\theta) \qquad (1)$$

as has been well established in the analyses of *Dst* based on a world-wide observation. In Equation (1), σ, a, r, e and i denote respectively the geomagnetic colatitude, Earth's radius, radial distance from the centre of the Earth, external and internal parts of the coefficient of the potential, while P_1 is the zonal spherical harmonic whose degree is equal to 1. The geomagnetic North and downward components are then given as

$$\left.\begin{aligned} \Delta X_m &= -(e+i)\sin\theta, \\ \Delta Z &= (e-2i)\cos\theta \end{aligned}\right\}. \qquad (2)$$

From the data observed at the nine Japanese observatories at the maximum epoch of the *Dst*, $e+i$ and $e-2i$ can be determined by the method of least squares.

We then have

$$e+i = 131\gamma, \qquad e-2i = 55\gamma$$

from which it follows that

$$e = 105\gamma, \qquad i = 25\gamma, \qquad e/i = 4{\cdot}2.$$

Meanwhile, the ratio e/i of *Dst* at the maximum epoch of main phase has been obtained as 3 or thereabout from analyses (Chapman & Price 1930, Rikitake & Sato 1957) of world-wide data. Attention should be paid to the fact that the internal part is considerably smaller when the coefficients are determined from observations in Japan only. The fact suggests that the electrical state at a depth of several hundred kilometres beneath Japan is likely to differ from what we have presumed for the average state of the Earth. Underneath Japan, it may be presumed that either the conductivity below a depth of a few hundred kilometres is smaller than the mean value of the Earth as a whole or the depth of the weak conducting layer, which covers the high conducting part of the mantle, is larger than that for the mean Earth. Although it is difficult to determine which conclusion is more likely, the result seems to agree very well with the low conductivity beneath Japan as has been suggested from the analysis of *Sq*.

7. Possible cause of the anomalies

The anomaly of short-period geomagnetic variations as observed in Japan is hardly understandable from the standpoint of the theory of electromagnetic induction within a uniform Earth. Even if we suppose a conductor in the upper part of the mantle where the conductivity is believed to be usually very low, no magnetic field large enough for the explanation of the anomaly can be produced by the induced electric currents as long as the conductor is isolated from the conducting part of the mantle in farther depths (Rikitake 1959).

The only way to account for the anomaly would be to suppose a circuit, both ends of which are connected to the conducting part of the mantle below a depth of a few hundred kilometres. If the circuit forms a loop in a suitable way beneath Japan, a part of the electric currents induced in the mantle might flow along the loop and produce a fairly strong magnetic field near the centre of the loop. On the basis of a model experiment, Nagata & others (Nagata, Oguti & Maekawa 1955) have shown that there might be a possibility of explaining the anomaly by such a branch circuit.

Although it is not possible to conduct an exact study in the framework of the existing theory of electromagnetic induction, the amount of electric current which flows in the circuit at times of geomagnetic bays, can be estimated very roughly on the basis of a quasi-stationary treatment. It turns out, however, that even if we take the largest possible value, it is hardly acceptable that the current exceeds several hundred amperes for a bay, the external part of which amounts to 100γ, provided a circuit appropriate to account for the anomalous area is considered. In that case, a magnetic field amounting to only a few gammas is to be observed at the centre of the loop. The hypothesis of such a branch circuit must be therefore abandoned unless some other means is found to increase the electric potential difference between both the ends of the circuit.

One of the most straightforward ways to increase the potential difference is to suppose an insulating plane between the two connecting points of the branch circuit. A two-dimensional experiment of steady electric current shows that it is not impossible to make the potential difference 50 times larger than that without

the insulating cut. A detailed study (Rikitake 1959) shows, for example, that a 1-hour period inducing field of 100 gammas in amplitude may be able to induce electric currents amounting to 30 000 amperes in a circuit, the ends of the circuit being connected to the conducting part of the mantle at two points 1 000 km apart. The upper part of the circuit that forms a circular loop 100 km in radius is also assumed to be composed of a substance of which the electrical conductivity amounts to 10^{-10} emu, while the radius of the section of the circuit is assumed as 10 km. The cut reaches a depth of several hundred kilometres in this case. The magnetic field produced by such induced currents is large enough to cause the anomaly observed in Japan.

In the light of the above, we see that a rather artificial distribution of electrical conductivity is required in order to explain the anomaly of short-period variations in Japan. Although it seems difficult to conduct more accurate investigations, the best and easiest way to account for the anomaly is to suppose a high-conducting circuit which comes up from the deep forming a roughly elliptical loop at 100 ~ 200 km depth beneath Japan. It is also required to suppose a deep low-conducting wedge there.

The supposition of the low-conducting wedge harmonizes very well with the low conductivity in the mantle at a depth of several hundred kilometres which has been suggested from the analyses of *Sq* and *Dst*. If the substance which usually consists of the upper layer of the mantle penetrates as deep as 700 km beneath Japan, the anomalies for such slow variations may be explained.

Summarizing all the analyses related to the anomalies of geomagnetic variations in Japan, a crude picture of the underground structure is supposed in the following. Beneath Japan, the conductivity at depths of several hundred kilometres seems likely to be very low, probably something of the order of 10^{-15} emu in contrast to 10^{-12} emu for the average value of the whole Earth. The boundaries of the low-conducting region are not known, though it is supposed that the region is not so wide as to cover China or South-west Japan because no marked anomaly of *Sq* is observed in those districts. It is likely that the region spreads beneath the central and north-eastern parts of Japan, while no information has been obtained about the eastern boundary which probably lies under the Pacific Ocean.

Above the low-conducting region, there must be a highly conducting circuit of electric currents presumably at depths of 100 ~ 200 km. The circuit should be connected to the conducting part of the mantle extending outside the low-conducting region presumed in the above. Although no accurate estimate of the dimension of the circuit is possible, there seems to be no objection to assuming that the circuit forms a roughly elliptical loop in an approximately horizontal plane whereas its width and length amount to about 200 and 1 000 km respectively.

8. Geophysical significance of the presumed underground structure

The bearing of the extraordinary distribution of the electrical property on geophysical or geological problems is not clear. Rikitake (1956) has presented, however, a hypothetical interpretation in relation to the underground structure of the orogenic belt running along Japan.

It is assumed that the upper part of the mantle wedges deep into the Earth's mantle along the Kamchatka–Japan–Mariana arc with a certain width, a likely value of which is, say, 200 km. Outside the hypothetical region the conductivity is taken as the same as that obtained for the average state of the Earth, that is, it

amounts to 10^{-15} emu down to a depth of 400 km increasing steeply at this level to 10^{-12} emu.

As a result of a simple estimate based on the theory of ionic crystals, we see that the material which usually occupies the part of the mantle from its top to the 400 km deep would have a conductivity around 10^{-15} emu even under the pressure and temperature at the depth of 700 km. If it goes deeper, an enormous increase in the conductivity would take place. We therefore presume that the lower limit of the penetrating wedge would not exceed 700 km as has been already suggested from the analysis of the *Sq*.

If the wedge is once taken for granted, the penetrating part undergoes some buoyancy force because of the difference of densities in and outside the region. With a few assumptions on the physical properties of matter, the buoyancy force acting there can be estimated. If the dip-angle of the penetrating part is supposed to be 45° as one may imagine from the way of occurrences of deep-focus earthquakes, the shearing stresses in the dipping plane would be slightly less than the strength of the Earth. Such stresses might cause earthquakes along the plane. The fact that no earthquakes have their foci deeper than 700 km may be also compatible with the hypothesis.

As a consequence of the stresses constantly acting near the boundary of the supposed wedge, some fractures could be formed along the boundary. If so, magma may rise up from there. Though an idea of this sort is highly speculative, a number of geologists are of the opinion that some effusive rocks might come up from as low a depth as several hundred kilometres. The trend of the volcanic zones is also understandable provided we assume such fractures.

Turning to the highly conducting circuit that has been inferred from the anomaly of short-period geomagnetic variations, no satisfactory interpretation can be given. It might be possible to correlate the circuit, which seems to be composed of high temperature material, with the fractures supposed above, though we cannot find any direct evidence.

Finally, the supposed wedging of the upper mantle might be caused by a dragging force due to a convection in the mantle. This assumption is also favourable for constructing a deep trench associated with the orogenic belt. It is not necessary to assume that the convection is now prevailing. If we assume that some convection had occurred at some time in the past when the Earth had more fluidity, the supposed structure could be formed.

Since the discussion in the above is of highly speculative character, we should not put much stress on the hypothetical interpretation. It is of interest, however, that a subterranean structure which seems to be closely correlated with orogenesis, seismicity and so forth, is brought out from the studies of anomaly of geomagnetic variations in Japan.

In conclusion, the writer wishes to express his cordial thanks to Professor T. Nagata, Sir Edward Bullard, Professor A. T. Price and Professor J. Bartels, through discussions with whom the writer was greatly encouraged. The writer is also indebted to Dr I. Yokoyama, Dr S. Uyeda and Mr T. Yukutake for their discussion and help during the eight-year period that the present study has been in progress.

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The Distribution of Shearing Stresses in a Tidal Current

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Summary

Observations were made in a tidal current off Red Wharf Bay, Anglesey, North Wales. The frictional stress at the bottom, F_b, was determined from the velocity profile within the first 2 m above the bottom and found to be related to the velocity at 1 m by a quadratic law, $F_b = k\rho U_1^2$, where k has the value $3{\cdot}5 \times 10^{-3}$. The corresponding value of the roughness length z_0 is 0·16 cm. Current meter measurements at a number of depths between surface and bottom were made at half-hourly intervals, enabling the acceleration terms in the equations of motion to be determined. From the bottom stress and the acceleration terms, the shearing stress in the water was computed as a function of depth and as a function of time during the tidal period. While at the times of maximum current the shearing stress increased approximately linearly from surface to bottom, as in the case of steady flow in a channel, at other times the acceleration terms caused the stress to deviate considerably from a linear variation. Estimates of the vertical eddy viscosity, N_z, indicated that its value was somewhat higher at mid-depth than nearer the surface or bottom. N_z varied during the tidal period, tending to reach maximum values when the current was at a maximum and to be larger during the flood than during the ebb. The numerical values of N_z were of the order of 270 cm²/s during the flood and 130 cm²/s during the ebb, corresponding to depth-mean currents of 35 cm/s and 39 cm/s respectively. The depth of water averaged 22 m. The observed distributions of velocity and shearing stress are compared with those obtained from a theoretical model, in which the eddy viscosity is taken as constant above a friction layer near the bottom.

1. Introduction

In the steady flow of a liquid of uniform density along an open channel, with friction at the channel bed, it follows directly from dynamical considerations that the shearing stress in the liquid increases linearly from the surface to the bottom. This result is valid, whatever the state of turbulence within the fluid. If some assumption can be made about the structure of the turbulent flow, e.g. von Karman's similarity law, the distribution of velocity with depth can be determined theoretically, as discussed by Hunt (1954). The well-known logarithmic law for the velocity distribution near the bed follows as a special case, in the

region near enough to the bed for the variation of the stress with height to be inconsiderable (less than 10 per cent, say, of its value at the bed).

In the general case of the flow varying with time and when the effects of the Earth's rotation have to be taken into account, as in the case of tidal currents, no direct inference can be drawn about the distribution of shearing stress with depth. The influence of bottom friction on tidal currents has been the subject of a number of investigations; some, following Taylor (1918) and Jeffreys (1920), based on the dissipation of energy over a large area of sea bed and others on detailed observations in a small area. In an investigation by the latter method (Bowden & Fairbairn 1952), the observations were treated by harmonic analysis and some indication found that, assuming the presence of a harmonic constituent of shearing stress corresponding to that of the velocity, the amplitude of the stress increased linearly with depth. This implies a certain degree of similarity between the stress distribution in steady flow and in a tidal current, but it does not show how the stress varies in the course of a tidal cycle. Since the bottom stress varies non-linearly with velocity and the turbulence in the interior of the liquid may also be expected to have a non-linear variation, it seemed desirable to make observations which would disclose the variation of the shearing stresses within a tidal cycle.

The method of Bowden & Fairbairn (1952) involved measuring the surface gradients by pressure gauges laid on the sea bed, and the accuracy attainable was such that significant results could be obtained only by the harmonic analysis of observations extending over a 24-hour period.

The aim of the present investigation was to measure the vertical distribution of current throughout the tidal cycle with sufficient accuracy to enable the acceleration terms in the equations of motion to be computed, and at the same time to estimate the bottom friction independently. From these data the shearing stress may be determined as a function of depth at various times during the tidal period.

2. Theory

Let rectangular co-ordinates x, y, z be taken, with the origin O in the mean level of the free surface and Oz vertically upwards. Let t denote time, u, v the components of velocity in the x, y directions, ζ the elevation of the free surface, ρ the density of the water, g the acceleration due to gravity, ω the angular speed of the Earth's rotation and ϕ the latitude. Let F_{zx}, F_{zy} be the components of the frictional force per unit area by which the water above a horizontal plane at depth z acts on the water below that plane. It is assumed that the shearing stresses acting on vertical planes are negligibly small.

Let it be assumed that the direct effect of the tide-generating force and the vertical components of velocity and acceleration may be neglected, and that conditions are sufficiently uniform in the horizontal directions for terms such as $u\partial u/\partial x$ to be neglected.

If it is assumed, also, that the density of the water is uniform, then the equations of motion for an element of water at any point may be written

$$\frac{\partial u}{\partial t} - fv = -g\frac{\partial \zeta}{\partial x} + \frac{1}{\rho}\frac{\partial F_{zx}}{\partial z} \tag{1}$$

$$\frac{\partial v}{\partial t} + fu = -g\frac{\partial \zeta}{\partial y} + \frac{1}{\rho}\frac{F_{zy}}{\partial z} \tag{2}$$

where $f = 2\omega \sin\phi$.

When the acceleration terms on the left-hand sides of these equations are zero, it follows that F_{zx} and F_{zy} are linear functions of z.

In the general case, integrating from the surface $z = 0$ to the bottom $z = -h$, and putting

$$\bar{u} = \frac{1}{h}\int_{-h}^{0} u\,dz, \quad \bar{v} = \frac{1}{h}\int_{-h}^{0} v\,dz,$$

$F_{zx} = F_{sx}$ and $F_{zy} = F_{sy}$ at $z = 0$,
$F_{zx} = F_{bx}$ and $F_{zy} = F_{by}$ at $z = -h$,
then

$$\frac{\partial \bar{u}}{\partial t} - f\bar{v} = -g\frac{\partial \zeta}{\partial x} - \frac{F_{bx} - F_{sx}}{\rho h} \tag{3}$$

$$\frac{\partial \bar{v}}{\partial t} + f\bar{u} = -g\frac{\partial \zeta}{\partial y} - \frac{F_{by} - F_{sy}}{\rho h}. \tag{4}$$

Hence from (1) and (3) and from (2) and (4) respectively,

$$\frac{\partial}{\partial t}(u - \bar{u}) - f(v - \bar{v}) = \frac{1}{\rho}\frac{\partial F_{zx}}{\partial z} + \frac{F_{bx} - F_{sx}}{\rho h}, \tag{5}$$

$$\frac{\partial}{\partial t}(v - \bar{v}) + f(u - \bar{u}) = \frac{1}{\rho}\frac{\partial F_{zy}}{\partial z} + \frac{F_{by} - F_{sy}}{\rho h}. \tag{6}$$

Writing these equations in finite difference form with $\Delta_t u$ etc. denoting the increment in u etc. in a time interval Δt, (3) gives

$$G_x \equiv -g\frac{\partial \zeta}{\partial x} = \frac{\Delta_t \bar{u}}{\Delta t} - f\bar{v} + \frac{F_{bx} - F_{sx}}{\rho h} \tag{7}$$

and (4) gives

$$G_y \equiv -g\frac{\partial \zeta}{\partial y} = \frac{\Delta_t \bar{v}}{\Delta t} + f\bar{u} + \frac{F_{by} - F_{sy}}{\rho h}. \tag{8}$$

Thus the surface gradients may be computed in terms of the mean current and the stress at the surface and the bottom. From (5) and (6),

$$-\frac{1}{\rho}\frac{\partial F_{zx}}{\partial z} = \frac{F_{bx} - F_{sx}}{\rho h} - A_x \tag{9}$$

$$-\frac{1}{\rho}\frac{\partial F_{zy}}{\partial z} = \frac{F_{by} - F_{sy}}{\rho h} - A_y, \tag{10}$$

where

$$A_x = \frac{\Delta_t(u - \bar{u})}{\Delta t} - f(v - \bar{v}) \tag{11}$$

and

$$A_y = \frac{\Delta_t(v - \bar{v})}{\Delta t} + f(u - \bar{u}). \tag{12}$$

Integrating (9) with respect to z,

$$F_{zx} = F_{sx}\left(1+\frac{z}{h}\right) - F_{bx}\frac{z}{h} + \rho\int_0^z A_x\,dz, \tag{13}$$

where a constant is added so that $F_{zx} = F_{sx}$ at $z = 0$. Integrating (10) gives a similar equation for F_{zy}.

If the surface stress can be neglected, so that $F_{sx} = F_{sy} = 0$, and the integration is carried out by finite intervals Δz, (13) becomes

$$F_{zx} = -F_{bx}\frac{z}{h} + \rho\sum_{z=0}^{z} A_x\Delta z. \tag{14}$$

Similarly

$$F_{zy} = -F_{by}\frac{z}{h} + \rho\sum_{z=0}^{z} A_y\Delta z. \tag{15}$$

In this case the components of shearing stress F_{zx} and F_{zy} can be computed from a knowledge of (a) the bottom stress, with components F_{bx} and F_{by} and (b) the acceleration terms A_x, A_y, obtained from (11) and (12), using observations of the current, with components u and v, as a function of time at various depths.

If a coefficient of eddy viscosity N_z is defined so that

$$F_{zx} = \rho N_z\frac{\partial u}{\partial z}, \qquad F_{zy} = \rho N_z\frac{\partial v}{\partial z} \tag{16}$$

then, having determined F_{zx} and F_{zy}, N_z may be computed from

$$N_z = \frac{F_{zx}}{\rho}\left(\frac{\Delta_z u}{\Delta z}\right)^{-1}, \quad \text{or} \quad N_z = \frac{F_{zy}}{\rho}\left(\frac{\Delta_z v}{\Delta z}\right)^{-1}. \tag{17}$$

For the area of observations in Red Wharf Bay, $\phi = 53°21'$, so that $f = 1{\cdot}17\times10^{-4}\,\mathrm{s}^{-1}$ and taking the time interval Δt as 1 hour, $1/\Delta t = 2{\cdot}78\times10^{-4}\,\mathrm{s}^{-1}$, so that

$$A_x = [2{\cdot}78\Delta_t(u-\bar{u}) - 1{\cdot}17(v-\bar{v})]10^{-4}\,\mathrm{cm/s^2} \tag{18}$$

$$A_y = [2{\cdot}78\Delta_t(v-\bar{v}) + 1{\cdot}17(u-\bar{u})]10^{-4}\,\mathrm{cm/s^2} \tag{19}$$

for the values of A_x and A_y at a time t hours, the increment Δt being taken from $t-\frac{1}{2}$ to $t+\frac{1}{2}$ hours.

3. Observations

Current measurements were made from an anchored vessel, the m.v. *William Herdman*, at three stations off Red Wharf Bay, Anglesey, North Wales, between 1957 July 16 and July 23. The positions of the stations and the depths at low water were:

Station 1: 53°19·6′ N, 4°6·0′ W, depth 12 m
Station 2: 53°21·2′ N, 4°7·0′ W, depth 16 m
Station 3: 53°22·3′ N, 4°11·2′ W, depth 19 m

At each station, bottom samples established that the sea bed consisted of firm sand with some shingle. Water samples, taken by insulating water bottle, showed that there was no appreciable variation in temperature or salinity from surface to bottom. Observations of bottom friction and turbulent flow had been made previously in the same area (Bowden & Fairbairn 1952, 1956).

The measurements of the water velocity from 4 m below the surface to 1 m above the bottom were made with a Doodson electrically recording current meter. This instrument, which has been described by Doodson (1940), was modified by the substitution of a continuously recording milliammeter with a paper speed of 1 in/min for the original direct-reading galvanometer or photographic recorder. The friction of the pen on the recording paper tended to smooth out some of the more rapid current fluctuations and so rendered the instrument unsuitable for the measurement of short-period turbulence. However, where the mean currents over a rather longer period were required, as in the present investigation, it was more convenient. A clockwork timing mechanism was also fitted to give an automatic changeover from direction to speed measurement and enabled a 5 s record of direction and a 25 s record of speed to be made every 30 s.

The direction readings are estimated to be correct to about 5° and except for low speeds (< 12 cm/s), where frictional effects in the mechanical system of the meter caused some uncertainty, the speed readings are regarded as correct to about ± 1 cm/s.

In addition, the mean water velocity in the 2 m immediately above the sea bed was measured at five fixed distances from the bottom. The instrument used for this purpose was developed by Charnock (1959) and consisted of five sets of cup-wheels mounted on a steel tube of approximate length 2·2 m. The tube itself was rigidly attached to a tripod base, which was specially designed to give firm seating on the sea bed. Each cup-wheel consisted of a Tufnol housing containing monel metal races and phosphor bronze balls, and was fitted with three equally spaced aluminium anemometer cups. Two small magnets were fixed diametrically opposite to one another on the sides of the rotating part of the housing, and, in the course of their motion, operated a magnetic switch set immediately below the cup-wheel. The switches were each connected to a GPO type counter situated aboard ship; thus the rate of counting gave a measure of the water velocity. Counts were continued over a period of 5 min and, in the absence of any obstruction to the rotation of the wheels, it is estimated that mean velocities were obtained to ± 1 cm/s. Some trouble was experienced during the course of the experiments due to floating weed and jelly-fish, but it was possible to detect and discard spurious readings because of their significant difference from those given by other instruments.

4. Analysis of velocity profiles

The Doodson meter records of the speed and direction of the current were made at intervals of 4 m from 4 m below the surface downwards, the lowest record being taken at 1 m above the bottom. Each record lasted for 3 min and a record at each depth was obtained every $\frac{1}{2}$ hour.

The readings of the records were converted into velocity in cm/s and direction in degrees relative to magnetic north, from the calibration of the meter. For each station the predominant direction of the current was estimated and this direction

taken as defining the x-axis, positive in the sense of the flood current, for that station. The individual current readings were then expressed in u and v components, which were plotted against time for each depth and smooth curves were drawn through the points. From these curves the smoothed values of u and v were read off and plotted against depth for each $\frac{1}{2}$-hourly interval of time relative to high water (HW). From the resulting depth-profiles of u and v, the values at standard values of z/h, i.e. $-z/h = 0{\cdot}25$, $0{\cdot}5$, $0{\cdot}75$ and $0{\cdot}95$, were read off and comprised the data for computing the accelerations.

Measurements of mean velocity were made with the cup-wheel device over a 5 min period every 30 min. The wheels were situated at distances of $0{\cdot}3$, $0{\cdot}6$, $1{\cdot}0$, $1{\cdot}5$ and $2{\cdot}0$ m from the seabed. Prior to use in the sea, the instruments had been calibrated in a water channel, thus it was possible to convert the counter readings directly into cm/s. Graphs of mean current against time were drawn for each height above the seabed. Spurious values of current, due to instrumental faults, were detected very easily by comparison of the curves and were omitted in the subsequent evaluation of bottom stress.

5. Determination of bottom stress

It was assumed that the velocity profile near the sea bottom was logarithmic in form and could be represented by

$$U_z = \frac{1}{k_0}\left(\frac{F_b}{\rho}\right)^{\frac{1}{2}} \ln\frac{z}{z_0}.$$

In the above formula, U_z is the mean velocity at distance z from the bottom, k_0 is von Karman's constant with value $0{\cdot}4$, F_b is the bottom stress, ρ is the density of the water and z_0 is the roughness length.

The groups of simultaneous values of velocity at different levels which contained no obviously spurious measurements were now fitted to the logarithmic formula, using the method of least squares. Values of F_b were then deduced and graphs drawn showing the variation of bottom stress with time. An example of these graphs is shown in Figure 1. Some of the calculated values of stress cause the curve to exhibit marked divergences from the relatively smooth curve which might be anticipated with a tidal current. In some measure, this was certainly due to errors in the determination of the mean currents, since small errors in

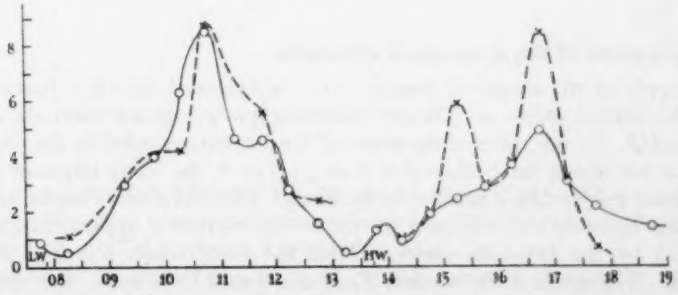

FIG. 1.—Bottom stress, F_b, as a function of time, at Station 2 on 1957 July 17: ——×—— from logarithmic profiles; ——○—— from $k\rho U_1^2$.
Ordinate: F_b(dyn/cm^2); abscissa: hours, G.M.T.

these values were capable of producing considerable errors in stress because of the logarithmic relation. It is believed, however, that turbulent fluctuations of current involving periods of the order of 1 min or more were partly responsible and further investigation of the effect is in progress.

The tripod stand could not be used when the ship was swinging about its anchor and thus there were gaps in most of the stress variation curves. There were additional intervals due to deficiencies in some of the current data. In order to fulfil the major purpose of the investigation it was essential to employ a method which would enable estimates of stress to be made throughout the tidal period. The mean current graphs at 1 m from the bottom were more complete than those for stress and it was possible to fill most of the gaps by interpolation. In terms of the mean current U_1, at 1 m from the bottom, the frictional force F_b was assumed to be of the form $F_b = k\rho U_1^2$, where k was a constant. Estimates of k were obtained from all cases where simultaneous values of the current at 1 m and the stress, as determined from the velocity profile, were available, and the mean value was found to be $3{\cdot}5 \times 10^{-3}$. This value of k was then used to find F_b for every satisfactory value of U_1 in the series of measurements. Figure 1 includes a curve showing the variation with time of stress deduced in this way, for comparison with the curve showing the direct determination of stress from the logarithmic profiles.

The value of $3{\cdot}5 \times 10^{-3}$ for k is in agreement with that found by the previous method (Bowden & Fairbairn 1952) in the same area. In that case a value $k' = 1{\cdot}8 \times 10^{-3}$ was derived, but k' was defined as relating the amplitude of a harmonic constituent of the stress to that of the mean current $\bar{U}$. In order to convert k' to k as defined above, i.e. referring the instantaneous stress to the current at 1 m, it is necessary to multiply by the factor $(3\pi/8) \cdot (\bar{U}/U_1)^2$. Since $U_1/\bar{U}$ was found to be 0·77, this factor is 2·0 and $k' = 1{\cdot}8 \times 10^{-3}$ corresponds to $k = 3{\cdot}6 \times 10^{-3}$. A lower value, $k = 2{\cdot}4 \times 10^{-3}$ was found by equating the stress at the bottom to the Reynolds stress determined directly from the turbulent velocity fluctuations at 75 cm (Bowden & Fairbairn 1956). Subsequent measurements have shown, however, that the mean currents were overestimated in those observations, and the corrected values of k range from $2{\cdot}5$ to $4{\cdot}4 \times 10^{-3}$, with a mean of $3{\cdot}5 \times 10^{-3}$, but the exact agreement is fortuitous.

The roughness length was also found in each velocity profile and the mean value of 0·16 cm is in agreement with those obtained in similar conditions by Lesser (1951) and Charnock (1959).

6. Application of the dynamical equations

In applying the theory of Section 2 to the observed data at a particular station, the bottom stress at ½-hourly intervals was computed from the equation $F_b = k\rho U_1^2$. U_1 was taken as the speed of the current recorded by the cup current meter at 1 m above the bottom and k as $3{\cdot}5 \times 10^{-3}$, the value obtained from the logarithmic profile data discussed in Section 5. The direction of the bottom stress was taken to be directly opposed to that of the current at approximately 1 m, as measured by the Doodson meter. Thus the components F_{bx} and F_{by} were obtained. The stress at the surface, F_s, was assumed to be zero. For most of the time the estimated wind speed W did not exceed 10 kt (about 5 m/s), for which the stress, computed from the equation $F_s = k_a \rho_a W^2$ with $k_a = 2{\cdot}5 \times 10^{-3}$, would be 0·8 dyn/cm^2. For certain periods during the observations on July 17, 22 and 23,

higher wind speeds, up to 15 kt (about 7 m/s), occurred, but as no instrumental measurements of W were made it was thought preferable not to attempt to make a correction for the wind stress effects. As seen from equation (13), the effect of F_s on the internal shearing stress decreases with increasing depth below the surface.

The depth-means of the current components, $\bar{u}$ and $\bar{v}$, were computed for each $\frac{1}{2}$-hourly set of readings and the values of $u-\bar{u}$ and $v-\bar{v}$ obtained for the standard values of z/h. The acceleration terms A_x and A_y were then computed, using equations (18) and (19). In general both the time-variation and the geostrophic terms made significant contributions to A_x and A_y, although for the larger values of A_x the term $\Delta_t(u-\bar{u})$ was predominant. The values of F_{zx} and F_{zy} were then computed for $-z/h = 0{\cdot}375$, $0{\cdot}625$ and $0{\cdot}875$, from equations (14) and (15).

The values of G_x and G_y were also computed, from equations (7) and (8), but are not discussed here as they are not relevant to the shearing stresses.

7. Errors in the stress determinations

If U is the resultant current at any depth, making an angle θ with the Ox direction, then $u = U\cos\theta$, $v = U\sin\theta$. For the current observations it is estimated that the standard errors of measurement, making some allowance for the smoothing process, are $\delta U = 1$ cm/s, $\delta\theta = 5°$. Since Ox is chosen to be in the dominant direction of the current, θ is small during most of the tidal cycle. Taking $U = 40$ cm/s, $\theta = 10°$ as typical values, the standard errors in u and v are $\delta u = 1{\cdot}2$ cm/s, $\delta v = 3{\cdot}6$ cm/s. Assuming the errors in individual values of $u-\bar{u}$ and $v-\bar{v}$ to be of the same magnitude as those in u and v, since errors in $\bar{u}$ and $\bar{v}$ will not affect the *variation* of shearing stress with depth, the errors in $\Delta_t(u-\bar{u})$ and $\Delta_t(v-\bar{v})$ may be taken as $\delta[\Delta_t(u-\bar{u})] = 1{\cdot}7$ cm/s, $\delta[\Delta_t(v-\bar{v})] = 5{\cdot}1$ cm/s. Then the standard errors in A_x and A_y are $\delta A_x = 7\times10^{-4}$ cm/s^2 and $\delta A_y = 14\times10^{-4}$ cm/s^2.

Now the deviation of the stress profile from linearity is seen from equations (14) and (15) to depend on the terms $\rho\Sigma A_x\Delta z$ and $\rho\Sigma A_y\Delta z$. The increments used for $-\Delta z/h$, going from surface to bottom, were $0{\cdot}375$, $0{\cdot}25$, $0{\cdot}25$ and $0{\cdot}125$ respectively, so that taking $h = 24$ m, $-\Delta z$ would be 9, 6, 6 and 3 m respectively. The values of A_x, A_y at the various depths are not independent, but are such that

$$\sum_{z=0}^{-h} A_x\Delta z = 0$$

and

$$\sum_{z=0}^{-h} A_y\Delta z = 0.$$

Thus the errors in $\rho\Sigma A_x\Delta z$ and $\rho\Sigma A_y\Delta z$ will be greatest at $-z/h = 0{\cdot}375$ and $0{\cdot}625$. From the above figures it is estimated that the standard value of these errors are $0{\cdot}6$ dyn/cm^2 in $\rho\Sigma A_x\Delta z$ and $1{\cdot}3$ dyn/cm^2 in $\rho\Sigma A_y\Delta z$.

The errors in the absolute values of F_{zx} and F_{zy} depend also on the errors in the estimate of the components of bottom stress, F_{bx} and F_{by}, and in assuming the surface stress components, F_{sx} and F_{sy} to be negligible, F_b is taken to be given by the equation $F_b = k\rho U_1^2$, with $k = 3{\cdot}5\times10^{-3}$, and the only error in F_b is

assumed to be that arising from the error in U_1. Then if $U_1 = 40$ cm/s and $\delta U_1 = 1$ cm/s, $\delta F_b = 0{\cdot}28$ dyn/cm^2. If the direction of F_b makes an angle θ with the x-axis, then with $\theta = 10°$, $\delta\theta = 5°$, the errors in F_{bx} and F_{by} in the above case would be $\delta F_{bx} = 0{\cdot}3$ dyn/cm^2 and $\delta F_{by} = 0{\cdot}5$ dyn/cm^2.

In view of the larger errors in F_{zy} than in F_{zx}, and since F_{zy} is usually small compared with F_{zx}, the computed values of F_{zy} are not regarded as significant enough to be presented, and only the results for F_{zx}, the main component of the stress, are discussed.

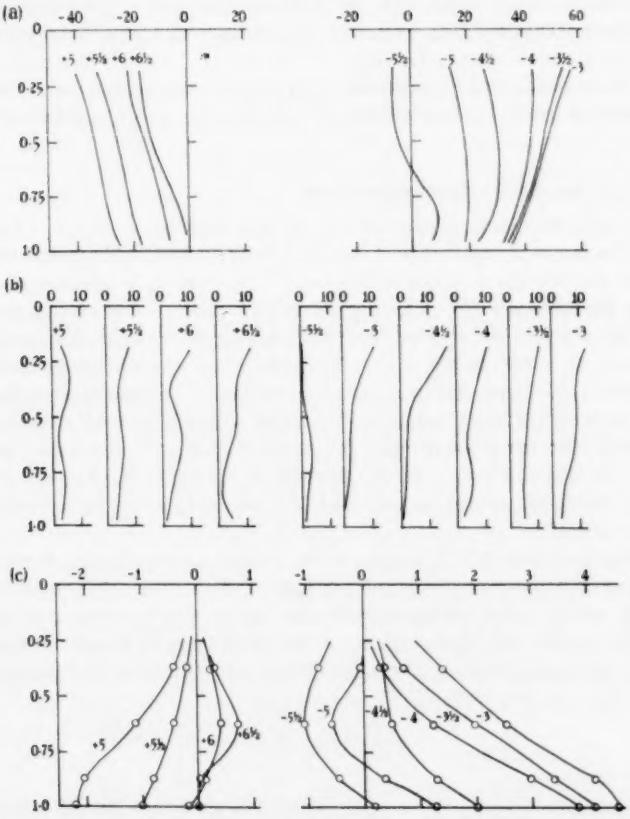

FIG. 2.—Curves of (a) u, (b) v, (c) F_{zx} as functions of z/h at half-hourly intervals. Station 2, 1957 July 19, 0730 to 1200 GMT (5 h after HW to 3 h before next HW).

Ordinate: (a) $-z/h$, (b) $-z/h$, (c) $-z/h$.
Abscissa: (a) u(cm/s), (b) v(cm/s), (c) F_{zx}(dyn/cm^2).

8. Results of the stress determinations

The results show that, although near maximum flood and maximum ebb currents the acceleration terms are small and the F_{zx} profile is nearly linear from surface to bottom, at other stages of the tidal cycle these terms have a considerable effect. The shape of the curve F_{zx} against z/h depends on whether the current is increasing or decreasing in magnitude and the curves show a systematic sequence

during the tidal cycle. When the current is increasing the curve is concave upwards, the stress at intermediate depths being less than that corresponding to a linear variation. When the current is decreasing, the effect is reversed, the stress at intermediate depths being greater.

Figures 2 and 3, referring to observations at station 2 on July 19, show a set of curves, at half-hourly intervals during a tidal period, of the current components u, v and the stress F_{zx}, computed as described above. Observations extending over a tidal period at station 2 were also obtained on July 17 and 23 and the results showed a similar sequence of curves for F_{zx}. At times on these two days, however, the wind stress had an appreciable effect on the velocity profile near the surface and hence on the computed stress at the smaller values of z/h.

The shapes of the F_{zx} curves indicate that F_{zx} at the smaller values of z/h lags behind the bottom stress F_{bx}. This effect is shown more explicitly in Figure 4, giving the variation of F_{zx} with time for the various values of z/h. In this figure the value of F_{zx} for each half-hour relative to HW represents a mean of the results for station 2, using the data obtained on the three days, July 17, 19, 23.

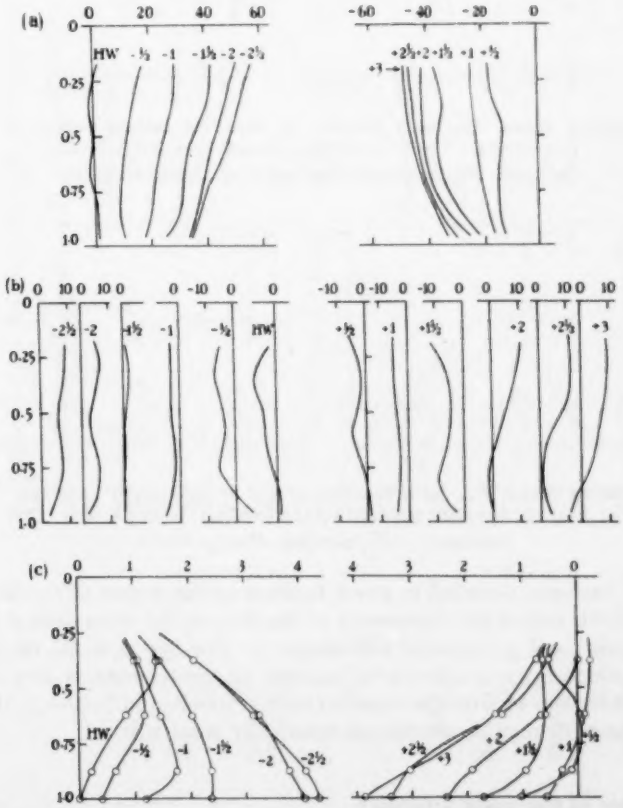

FIG. 3.—Curves of (a), u, (b) v, (c) F_{zx} as functions of z/h at half-hourly intervals. Station 2, 1957 July 19, 1230 to 1800 GMT (2½ h before HW to 3 h after HW).

Ordinate: (a) $-z/h$, (b) $-z/h$, (c) $-z/h$.
Abscissa: (a) u(cm/s), (b) v(cm/s), (c) F_{zx}(dyn/cm²).

Less complete results were derived for station 1 on July 16 and station 3 on July 22. The values of F_{zx} as a function of z/h at station 1 for a period of approximately 6 hours are shown in Figure 5. The amplitude of the current at 4 m on this occasion was 50 cm/s and the maximum bottom stress was 4·8 dyn/cm².

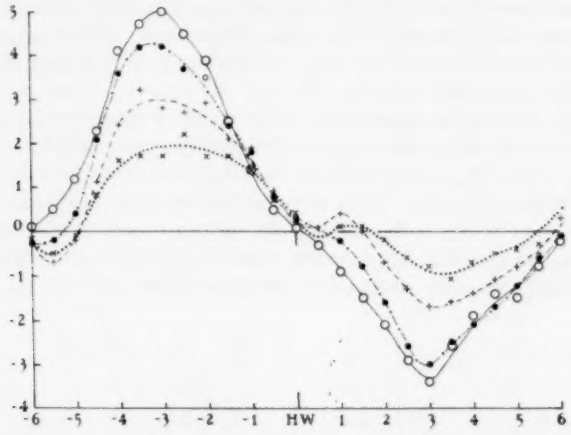

FIG. 4.—Shearing stress, F_{zx}, as a function of time, for various values of $-z/h$; i.e. · · ·×· · · 0·375, ——+—— 0·625, —·—●—·— 0·875, —○— 1.
Ordinate: F_{zx}(dyn/cm²); abscissa: hours relative to HW.

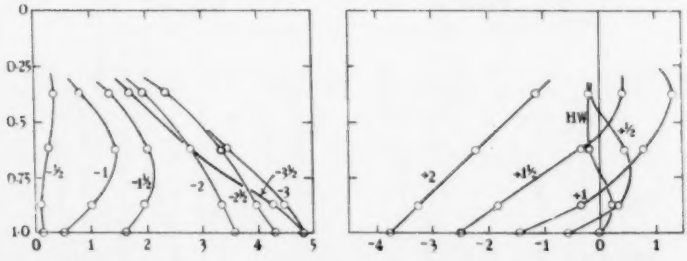

FIG. 5.—Shearing stress, F_{zx}, as a function of z/h at half-hourly intervals. Station 1 1957 July 16, 0939 to 1509 GMT (3½ h before HW to 2 h after HW).
Ordinate: $-z/h$; abscissa: F_{zx}(dyn/cm²).

Table 1 has been included to give a measure of the degree of non-linearity in the stress profile and of the consistency of the data on the three days at station 2, and at stations 1 and 3 compared with station 2. The figures in the table give the amount by which F_{zx} at $-z/h = 0{\cdot}625$ exceeds the stress corresponding to a linear profile. Taking into account the standard errors estimated in Section 7, the results shown in the table may be considered reasonably consistent.

9. Estimates of the eddy viscosity

Assuming that the stress at any depth at a given instant may be related to the corresponding velocity gradient by equation (16), the data enable N_z, the coefficient of eddy viscosity in the vertical direction, to be determined, using equation

Table 1

Deviation from linear stress profile, at $-z/h = 0{\cdot}625$

Deviation $\delta = F_{zx} - F_{bx}z/h$, in dyn/cm²

Station	Date	Hours before HW: 6	5½	5	4½	4	3½	3	2½	2	1½	1	½	HW
2	July 17	−0·8	−0·5	−0·6	−0·7	−0·2	0·9	−0·3	−0·5	0·2	0·5	1·6	0·1	0·4
	,, 19	0·7	−1·2	−1·4	−0·8	−1·1	−0·6	−0·3	0·5	0·7	0·6	0·7	0·9	0·9
	,, 23	−0·7	−1·2	−0·8	0·3	0·7	0·2	−0·5	—	—	—	0·8	0·7	−0·8
	Mean δ	−0·3	−1·0	−0·9	−0·3	−0·1	0·2	−0·4	0	0·5	0·5	1·0	0·6	0·2
	Mean F_{bx}	0·1	0·5	1·2	2·3	4·1	4·7	5·0	4·5	3·8	2·5	1·4	0·5	0·1
1	July 16	—	—	—	—	—	−0·2	0·4	0·7	0·6	1·0	1·1	0·1	−0·2
3	,, 22	0·1	−1·0	−1·1	−1·3	−0·3	0·1	0·6	1·7	2·3	0·7	−0·6	−0·4	0·5

Station	Date	Hours after HW: HW	½	1	1½	2	2½	3	3½	4	4½	5	5½	6
2	July 17	0·4	1·2	1·4	1·3	0·2	0·5	1·0	0·2	−0·6	−1·1	—	—	—
	,, 19	0·9	0·5	0·7	0·5	0·8	0·7	0·9	—	—	—	0·3	0·2	0·5
	,, 23	−0·8	−0·7	1·0	1·2	0·8	0·4	−0·4	−0·1	0·3	0·7	0	−0·4	−0·4
	Mean δ	0·2	0·3	1·0	1·0	0·6	0·5	0·5	0	−0·1	−0·1	0·1	−0·1	0
	Mean F_{bx}	0·1	−0·3	−0·9	−1·5	−2·1	−2·9	−3·4	−2·6	−2·0	−1·4	−1·5	−0·8	−0·2
1	July 16	−0·2	0·8	1·7	1·2	0·1	—	—	—	—	—	—	−0·2	−0·9
3	,, 22	0·5	0·7	0·9	—	—	—	—	—	—	—	—	—	—

Table 2

Values of eddy viscosity, N_z (cm²/s)

Station	Date	$-z/h$	Hours before HW										HW			Hours after HW							Mean
			5	4½	4	3½	3	2½	2	1½	1	½	0	½	1	1½	2	2½	3	3½	4	4½	5	
2	July 17	0·375	—	—	—	—	180	—	—	270	—	—	—	—	—	—	120	160	—	80	150	120	—	154
		0·625	—	—	350	610	—	360	—	430	—	—	—	—	—	—	150	260	120	120	130	160	—	269
		0·875	—	150	180	200	420	220	110	140	140	—	—	—	—	50	80	100	170	100	90	130	—	152
	,, 19	0·375	—	—	—	—	120	210	210	150	160	160	—	—	—	—	—	—	—	—	—	—	—	168
		0·625	—	—	170	220	250	205	390	310	220	250	—	—	—	—	—	200	140	—	—	—	150	228
		0·875	—	75	210	190	270	385	365	190	285	—	—	—	—	—	75	160	200	—	—	—	220	219
	,, 23	0·375	—	130	190	235	420	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	244
		0·625	—	150	410	520	210	—	—	—	—	—	—	—	—	—	—	—	150	90	—	—	—	250
		0·875	—	90	330	530	250	—	—	—	—	—	—	—	—	—	—	95	155	150	95	70	—	196
Mean		0·375	—	130	190	230	240	210	210	210	160	160	—	—	—	—	120	160	—	80	150	120	—	169
for		0·625	—	150	310	450	230	230	390	370	220	250	—	—	—	—	150	230	130	110	130	160	150	229
Stn. 2		0·875	—	110	240	310	310	300	240	170	210	—	—	—	—	50	80	120	170	130	90	100	220	178
		Mean	—	**130**	**250**	**330**	**260**	**250**	**280**	**250**	**200**	**200**				**50**	**120**	**170**	**150**	**110**	**120**	**120**	**180**	**192**
1	July 16	0·375	—	—	—	—	—	—	—	—	—	—	—	—	140	—	120	—	—	—	—	—	—	—
		0·625	—	—	—	215	280	280	365	270	—	—	—	—	—	—	140	—	—	—	—	—	—	260
		0·875	—	—	—	315	440	320	350	160	—	—	—	—	—	90	190	—	—	—	—	—	—	265
3	,, 22	0·375	—	—	—	—	70	160	250	—	—	—	—	—	—	—	—	—	—	—	—	—	—	160
		0·625	—	—	—	90	130	210	580	—	—	—	—	—	—	—	—	—	—	—	—	—	—	250
		0·875	—	—	—	65	95	130	220	150	—	—	—	—	—	—	—	—	—	—	—	—	—	130

Note: Values of N_z are given only for cases in which $|F_{zx}| \geqslant 1$ dyn/cm and $|\Delta_z u| \geqslant 3$ cm/s in 6 m.

(17). Because of the large errors in $\partial u/\partial z$ in addition to the errors in F_{zx}, the uncertainty in such estimates of N_z is bound to be large.

Table 2 contains estimates of N_z, computed for station 2 on the three days of observations. Only cases where $|F_{zx}| \geqslant 1$ dyn/cm and $|\Delta_z u| \geqslant 3$ cm/s for an increment of 6 m in Δ_z have been included. Even so, the estimated error in individual values of N_z is ± 50 per cent.

At first sight the values of N_z show no systematic variation with depth or time. However, by taking an average for each half-hour relative to HW from the data for the three days and then taking means (*a*) at the three values of z/h irrespective of time, and (*b*) for each half-hour relative to HW irrespective of depth, the following indications are found. (*a*) The eddy viscosity, N_z, tends to be higher near mid-depth (at $-z/h = 0{\cdot}625$) than either nearer the surface or nearer the bottom. (*b*) N_z tends to reach maxima about 3 hours before and 3 hours after HW, when the current is greatest. No reliable estimates of N_z are available at times close to HW, when the currents and stresses are small. There is also a tendency for higher values of N_z to occur during the flood than during the ebb, possibly associated with the higher velocities which occur during the flood. Taking the mean of all values of N_z, irrespective of depth, for times from 4 to 2 hours before HW (inclusive), one obtains 273 cm^2/s, while the similar mean for times from 2 to 4 hours after HW is 132 cm^2/s. The corresponding mean values of $\bar{u}$ are 44·7 and 39·4 cm/s, and of h are 22·0 and 22·5 m.

10. Comparison with a theoretical model

The method of analysis used in the foregoing paragraphs has involved no special assumptions about the nature of the turbulent flow. However, the comparatively small variation of N_z over a considerable part of the depth, shown by Table 2, combined with the approximately logarithmic form of the velocity profile nearer the bottom, indicated by the data discussed in Section 5, suggests that a simple theoretical model may provide an approximate representation of the flow. This model would consist of (*a*) a friction layer near the bottom, with a thickness of the order of one-tenth of the depth of water, in which the logarithmic velocity profile applies and N_z increases approximately linearly with height above the bottom, and (*b*) the remainder of the water, in which N_z is independent of depth. Similar models have been postulated by various authors in dealing with oceanographic and meteorological problems and the one proposed appears also to be consistent with the present state of knowledge of turbulent flow in channels on the laboratory scale (Townsend 1956, p. 202).

A theoretical solution for the tidal current under similar conditions was given by Sverdrup (1926) who, however, assumed the eddy viscosity to be constant down to the bottom and the velocity to be zero there. Street (1917) had treated a problem which was formally the same, although he was considering only the molecular viscosity. Fjeldstad (1929, 1936) obtained solutions with an eddy viscosity increasing with height above the bottom, but the bottom velocity was again taken to be zero.

In the present treatment the motion in the upper part of the flow only is considered, but allowance is made for the presence of the friction layer by imposing an approximate boundary condition at the upper limit of this layer. A harmonic constituent of the tidal current is considered and N_z is taken as independent of time. The basic equations for this problem are given in the Appendix, with the

solution for the case of flow near a straight coast, derived using an approximate form of the boundary condition. A numerical solution is obtained by taking values of the physical parameters corresponding to conditions in Red Wharf Bay.

The current component v, perpendicular to the coast, is shown to be small while the variation of u with z is similar to that found in the Red Wharf Bay area from the harmonic analysis of a number of observations (Bowden & Fairbairn 1952). Thus $|\bar{u}|/|u_s| = 0{\cdot}896$ compared with $0{\cdot}87$ from observations and $|u_b|/|u_s| = 0{\cdot}675$ compared with $0{\cdot}67$, taking u_b as the measured velocity at 1 m above the bottom. The phase of u_b is 3° earlier than that of u_s.

For the shearing stress F_{zx}, Table 3 shows that the theoretical amplitude increases with depth more slowly than a linear variation near the surface and more rapidly near the bottom. The phase becomes earlier with increasing depth to a considerably greater extent than does the phase of u. Thus at $-z/h = 0{\cdot}2$ the phase is 19° later than at the upper limit of the friction layer. Computed curves of F_{zx} against $-z/h$ for successive times, given in Figure 6, show a marked resemblance to the curves derived from observations in Figures 2, 3 and 5.

It appears therefore that, in spite of its obvious limitations, the simple model suggested may be a useful basis of calculation in problems of this nature.

Appendix

Flow of a tidal current, assuming a constant eddy viscosity above the friction layer

The notation of Section 2 will be used, except that $z = -h$ will be taken as the upper limit of the friction layer instead of the sea bottom. The same general conditions as in Section 2 will be assumed to apply. N_z is taken as constant in equation (16) for F_{zx} and F_{zy} and these expressions inserted in the equations of motion (1) and (2) of Section 2. Then it may be shown (as in Sverdrup 1926) that for a harmonic constituent of angular velocity σ, the velocity components are given by

$$u = F_1(x, y)U(z)e^{i\sigma t}, \qquad v = F_2(x, y)V(z)e^{i\sigma t}$$

where

$$\left.\begin{aligned} U &= 1 + A\cosh(1+i)\alpha_1 z + B\cosh(1+i)\alpha_2 z \\ V &= 1 - \frac{i}{c}\left[A\cosh(1+i)\alpha_1 z - B\cosh(1+i)\alpha_2 z\right] \end{aligned}\right\} \tag{A.1}$$

where

$$\alpha_1 = \left(\frac{\sigma+f}{2N_z}\right)^{\frac{1}{2}}, \qquad \alpha_2 = \left(\frac{\sigma-f}{2N_z}\right)^{\frac{1}{2}},$$

and A,B,C are constants to be determined.

At $z = -h$, the upper limit of the friction layer, the components of shearing stress are assumed to be given by

$$F_{bx} = k\rho|(u_b^2+v_b^2)^{\frac{1}{2}}|u_b, \qquad F_{by} = k\rho|(u_b^2+v_b^2)^{\frac{1}{2}}|v_b,$$

the suffix b denoting values taken at $z = -h$.

In the special case of flow near a straight coast parallel to Ox, this may be simplified to the approximate condition

$$N_z\left(\frac{dU}{dz}\right)_b = k'U_0U_b, \qquad N_z\left(\frac{dV}{dz}\right)_b = k'U_0V_b \tag{A.2}$$

where U_0 is the amplitude of the current in the absence of friction. k' is less than k by the factor $8/3\pi$ and by an approximate estimate of the expected ratio of the amplitude of $|(u_b^2+v_b^2)^{\frac{1}{2}}|$ to U_0.

Using (A.2) it is found that A, B and C are given by

$$A = -\frac{1+iC}{2P}, \qquad B = -\frac{1-iC}{2Q} \tag{A.3}$$

$$C = \frac{\dfrac{S_2}{2Q\alpha_2 h}-\dfrac{S_1}{2P\alpha_1 h}}{1-i\left(1-\dfrac{S_1}{2P\alpha_1 h}-\dfrac{S_2}{2Q\alpha_2 h}\right)} \tag{A.4}$$

where

$$\left.\begin{aligned} S_1 &= \sinh(1+i)\alpha_1 h, & S_2 &= \sinh(1+i)\alpha_2 h \\ C_1 &= \cosh(1+i)\alpha_1 h, & C_2 &= \cosh(1+i)\alpha_2 h \\ \text{and} \quad P &= C_1+\frac{(1+i)N_z}{k'U_0}\alpha_1 S_1, & Q &= C_2+\frac{(1+i)N_z}{k'U_0}\alpha_2 S_2 \end{aligned}\right\} \tag{A.5}$$

Then the solution for U and V is obtained by using these values of A, B and C in (A.1).

Numerical solution

Numerical values which represent approximately the conditions in Red Wharf Bay will be taken.

For the M_2 constituent $\sigma = 1{\cdot}407\times10^{-4}\,\mathrm{s}^{-1}$. For latitude 53°20′ N, $f = 1{\cdot}169\times10^{-4}\,\mathrm{s}^{-1}$. Let $h = 20\,\mathrm{m}$, $N_z = 250\,\mathrm{cm^2/s}$, $k' = 2\times10^{-3}$, $U_0 = 62{\cdot}5\,\mathrm{cm/s}$.

Then $\alpha_1 h = 1{\cdot}436$, $\alpha_2 h = 0{\cdot}436$, $k'hU_0/N_z = 1{\cdot}0$.

$S_1 = 2{\cdot}217\ \ (83{\cdot}1°)!$, $S_2 = 0{\cdot}617\ \ (48{\cdot}7°)!$, $C_1 = 1{\cdot}988\ \ (81{\cdot}3°)!$, $C_2 = 1{\cdot}012\ \ (10{\cdot}9°)!$.

Hence $A = 0{\cdot}057\ (102{\cdot}5°)!$, $B = 0{\cdot}667\ (133{\cdot}8°)!$, $C = 0{\cdot}608\ (46{\cdot}9°)!$.

It is found that $F_1/U_0 = 1{\cdot}390\ (-28{\cdot}6°)!$, $F_2/U_0 = 0{\cdot}845\ (18{\cdot}3°)!$.
At the surface, $F_1U_s/U_0 = 1{\cdot}045\ (17{\cdot}0°)!$, $F_1V_s/U_0 = 0{\cdot}018\ (196°)!$.
At the upper limit of the friction layer

$$F_1U_b/U_0 = 0{\cdot}705\ (20{\cdot}0°)!,\ F_1V/U_0 = 0{\cdot}029\ (348°)!.$$

The depth-mean, $F_1\bar{U}/U_0 = 0{\cdot}935\ (18{\cdot}4°)!$.
Thus $\bar{U}/U_s = 0{\cdot}896\ (1{\cdot}4°)!$ and $U_b/U_s = 0{\cdot}675\ (3{\cdot}0°)!$.

The stress component F_{zx} is given by the real part of

$$F_{zx} = \rho N_z \frac{\partial u}{\partial z} = \frac{(1+i)\rho N_z}{h} F_1 e^{i\sigma t}[A\alpha_1 h \sinh(1+i)\alpha_1 z + B\alpha_2 h \sinh(1+i)\alpha_2 z].$$

If this is written in the form

$$F_{zx}/(F_{bx})_{\max} = R_z \cos(\sigma t - \gamma_z)$$

the values of R_z and γ_z are as given in Table 3.

Table 3

Relative amplitude and phase of shearing stress (theoretical)

$-z/h$	R_z	γ_z (deg)
0	0	—
0·2	0·186	19·4
0·375	0·351	17·3
0·5	0·471	15·1
0·625	0·594	13·1
0·875	0·859	4·7
1	1	0

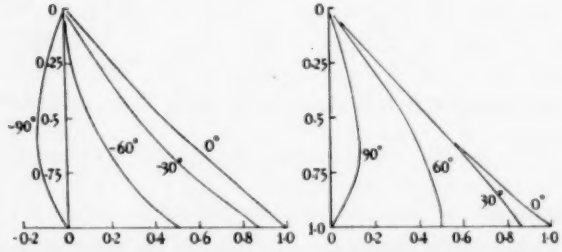

FIG. 6.—Theoretical curves of $F_{zx}/(F_{bx})_{\max}$ as a function of z/h at various phases of the tidal period, from $-90°$ to $+90°$ (corresponding to lunar hourly intervals from LW to HW).

Ordinate: $-z/h$; abscissa: $F_{zx}/(F_{bx})_{\max}$.

In Figure 6 curves are given for $F_{zx}/(F_{bx})_{\max}$ against $-z/h$, computed for values of σt from $-90°$ to $+90°$, corresponding to the time interval from 3 lunar hours before to 3 lunar hours after the occurrence of maximum shearing stress near the bottom.

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Palaeomagnetism of the Great Whin Sill

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Summary

The Great Whin Sill, which was intruded in Upper Carboniferous to Lower Permian times, has been sampled at 36 sites in Northumberland and Durham. It was found to be magnetically stable, the site mean direction of magnetization being 187°·8–4°·9 and the mean pole position lying in latitude 37°·3 N longitude 168°·9 E. The distribution of the mean directions of magnetization at these sites is not circular but oval, with the variation in the direction of the palaeomagnetic longitude about twice that in the direction of palaeomagnetic latitudes. However the poles calculated from the site mean directions of magnetization have a circular distribution, and this observation leads to proposal of a method of calculating the errors associated with the palaeomagnetic pole positions.

1. Introduction

The objects of this investigation were to determine the mean direction of magnetization of the Great Whin Sill and to investigate the pattern of secular variation. It has been argued (Watson & Irving 1957) that random errors are averaged out in the calculation of site-mean directions of magnetization, the latter representing spot readings of the geomagnetic field in an igneous body such as the Whin Sill. The distribution of the site mean directions of magnetization about their mean should therefore give the pattern of secular variation at the latitude in which the formation cooled through its Curie temperatures. As the Whin Sill is a composite intrusion and fairly thick, it seems probable that its cooling history would cover a time range sufficiently long such that its remanent magnetization would record the full range of secular variation.

In the statistical analysis of most previous work in palaeomagnetism it has been implicitly assumed that the directions of magnetization within a formation are circularly distributed around the mean direction, as proposed by Fisher (1953). Cones of confidence of half angle α within which the true mean directions of magnetization lie at given probabilities have been derived. The mean pole positions for the times represented by these formations have been computed from the mean directions of magnetization and areas of confidence have been constructed around them (Creer & others 1957). For formations magnetized at the poles these areas are circular in shape and of radius 2α. At the equator, they are ellipses with axes of half-length α and $\frac{1}{2}\alpha$, the latter in the direction of co-latitude. Between

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these two extremes of latitude their form is intermediate. Thus the shape of the areas of confidence so calculated depends on the palaeo-latitude. However, if we assume a circular distribution of the pole positions calculated for each site within a formation about the mean pole position for the formation as a whole, the observed directions of magnetization should have an oval distribution, this becoming more marked at low latitudes. Intense sampling of the Whin Sill was carried out to determine whether the mean directions of magnetization of sites or the mean pole positions of sites have circular distributions. This formation was chosen because earlier work by Hallimond & Butler (1949), and Manley (1949), had shown it to be almost horizontally magnetized. Furthermore the geology is well known and many natural and artificial outcrops are easily accessible.

2. The geology of the Whin Sill

The Great Whin Sill is the name given to a system of transgressive quartz dolerite sheets intruded into the Carboniferous rocks of northern England between latitudes 54°·6 N and 55°·6 N, and longitudes 1°·6 W and 2°·5 W (Figure 1). The outcrops sweep across Northumberland southward into County Durham, forming crags which, near Haltwhistle, carry the Roman Wall. It is also exposed in the west-facing Pennine escarpment. Associated with the Whin Sill is a system of petrologically similar dykes which may in part have been feeder dykes. The average thickness of the Whin Sill where it occurs as a single sheet is from 80 to 100 ft (Woolacott 1923), but as variation is both rapid and proportionately great, average thickness has little meaning.

The age of the intrusive system can be found within fairly close limits. The sill is generally intruded into lower Carboniferous rocks but associated dykes have been found cutting the Middle Coal Measures (Westphalian). In north Northumberland, at least, intrusion post-dates the folding of the lower Carboniferous beds (Westoll & others 1955). The upper limit, suggested by the fact that no dykes are known to cut Permian rocks, is better established by the discovery of rare pebbles of Whin type in the Upper Brockram of the Eden Valley. It can also be shown that intrusion preceded the mineralization of northern England (Holmes & Harwood 1928) and since detrital heavy mineral grains from the latter source have been found in the Permian Yellow Sands, the upper age limit can be regarded as reasonably fixed. In zonal terms the intrusion must be post-Westphalian pre-Thuringian, that is either Stephanian, Autunian or Saxonion, and Holmes and Harwood conclude "The interjection of the Whin Sill and related dykes may thus be regarded as the closing act of the Upper Carboniferous in the north of England". As will appear later in this paper, there are reasons for suggesting that an early Permian age is equally likely.

The Whin is a hard blue rock very much in demand as a road metal. Petrologically, the dominant Whin type is a medium-grained quartz dolerite. The occurrence of a coarse-grained dolerite pegmatite found as flattened lenticular masses in the upper parts of thick outcrops appears to be restricted to the south of the Tyne (Tomkeieff 1928). At contacts, a fine-grained marginal selvage, or occasionally vesicular, or even "white" Whin may be developed. Contacts of Whin against Whin, evidence of multiple intrusion, are known. Whin inclusions in Whin have also been described by Smythe (1914) as further evidence of multiple injection.

The average grain size of the normal type ranges from 0·2 to 1 mm. Microscopically, the structure is that of an interlacing network of plagioclase laths and

elongate or stumpy prisms of augite held together by plates or granular aggregates of augite with iron ore minerals in skeletal or well-individualized form. Ophitic, intergranular and intersertal textures are all well developed, frequently within a single section. Minor amounts of alkali felspar, quartz and micro-pegmatite occur interstitially with hornblende, biotite and accessory apatite. Chloritic and serpentinous minerals are the common alteration products. The sills are characterized by their homogeneity, the range of variation is so small that the diversity at a single exposure is comparable to that of the whole extent of the sill (Tomkeieff 1928).

From the magnetic point of view the important factors are the amount and composition of the opaque ores for which, on the whole, information is less complete. Micrometric measurements indicate that the opaque ores range from 5 to 10 per cent of the rock by volume, although according to Holmes & Harwood (1928) these estimates may be in error by as much as 50 per cent, particularly in the finer grained varieties. The normative magnetite calculated from chemical analyses averages from 4 to 5 per cent, while normative ilmenite is about 4 per cent. Akimoto (1955) has studied the ferromagnetic minerals in a typical specimen of Whin Sill and gives the cell dimension $a = 8{\cdot}415$Å (pure magnetite $a = 8{\cdot}413$Å) and a single Curie point at 565 °C (pure magnetite 575 °C) from which it appears that the magnetic mineral is very close to pure magnetite. If ilmenite occurs it must be as a separate paramagnetic phase. Pyrrohitite has only been recorded around inclusions of Whin materials within a Whin dyke (Smythe 1914).

The Whin intrusions have produced thermal metamorphic minerals in narrow zones at the contacts. Recrystallization occurs in some limestones while sandstones may locally become quartzites, and hornfelses may even develop. Garnet, wollastonite, idocrase and andalusite have all been found.

3. Method of sampling and measurement

Samples of the dolerite have been taken at 36 sites (quarries, sea cliffs or inland scarps) distributed along the outcrop (Figure 1). Three samples were obtained from each site, great care being taken to select fresh material. An exposed face of each was oriented by standard geological methods before extraction. A single core $1\frac{3}{8}$ in. in diameter and height with its axis perpendicular to the orientated face was machined from each sample using non-magnetic tools. The directions of magnetization of the cores were measured by a short period astatic magnetometer of the type described by Collinson & others (1957). The overall accuracy of orientation and measurement is 5°. Errors from this source are random, since the orientated surfaces were selected at random, and are therefore eliminated when the average direction of several specimens is taken. Of the 108 samples collected, three from a single site (No. 34) gave random directions and were rejected, while three others distributed among three other sites broke up during machining and were not used.

At three sites (16, 17 and 19), several samples of sedimentary rocks were collected from near the lower contact. From each of these samples a single one-inch core was cut and then sliced into several disks. The directions of magnetization of the disks were measured by a sensitive astatic magnetometer of the type described by Collinson & others (1957).

FIG. 1.—Map showing the distribution of sampling sites along the outcrop of the Whin Sill of north-eastern England.

4. The magnetization of the dolerite

At any one place the sills are either horizontal or dip at low angles, generally less than 10°. As it is uncertain whether tilting everywhere preceded or post-dated injection, the directions of magnetization have been worked out in two ways:

(a) with the beds in their present position, the assumption being that injection occurred after folding;

(b) with the sills everywhere reduced to a horizontal plane on the assumption that tilting occurred after intrusion.

The directions of magnetization in the 102 cores of dolerite are given according to the two methods in Figures 2a and 2b respectively. The plots are on stereographic equatorial projections so that positive and negative dips do not obscure one another and the shape of the distribution can be more clearly seen. The directions are consistently southerly with low dips and make angles of over 90° with the direction of the present field. Statistical tests as described by Watson & Irving (1957) show that the two distributions (a) and (b) are not Fisherian. The dispersion appears to decrease very slightly after correction for dip. The mean site directions, corrected to a horizontal plane, are plotted in Figure 3 with the overall

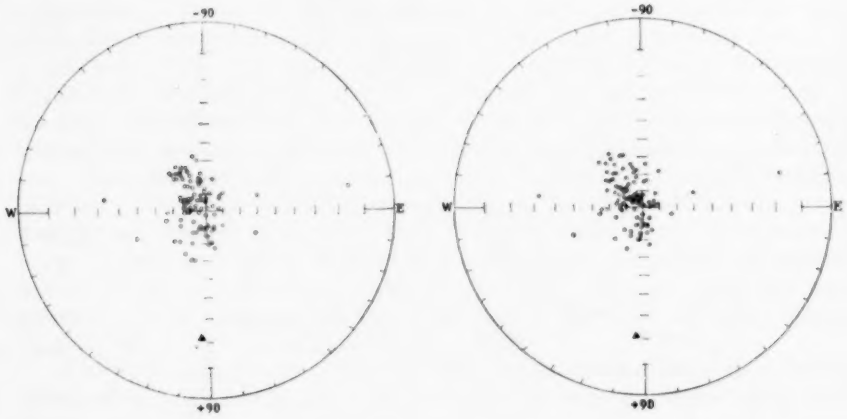

FIG. 2.—Stereographic equatorial plots of the directions of magnetization of Whin Sill samples.

(a) Results plotted with respect to the present field position.
(b) Results plotted with the Whin Sill restored to a horizontal plane.

mean direction at the centre of the stereogram. For both site and specimen directions the distribution of individual values around the mean direction is oval not circular with the variation in inclination about twice that in declination.

Certain magnetic properties of samples from five sites chosen at random were measured. Samples were placed in a d.c. magnetic field and the remanent magnetization (i.r.m.) on removal from the field was measured. The strength of the d.c. field was increased in steps and it was found possible to saturate the samples. The maximum i.r.m.s so obtained for samples from the five sites are recorded in Table 2. Next the samples were placed in d.c. magnetic fields directed in opposition to the maximum i.r.m. so obtained. The reduced intensity of remanent magnetization was measured for successively higher d.c. field strengths. The field strength required to remove the maximum i.r.m. is here defined as the coercivity of remanence and is also recorded in Table 2. The values of the two above quantities are similar to those previously given for basic igneous rocks (Nagata 1953, p. 107 and Creer 1955) and indicate that the magnetic minerals present are predominantly of the magnetite rather than the haematite family.

The initial susceptibilities (in d.c. fields of the order of the Earth's field) of samples from the same five sites were also measured. The values, also recorded

Table 1

Directions of magnetization of the Whin Sill

Site No.	Locality	National Grid reference	Site Directions of Magnetization (a) referred to present horizontal		(b) corrected to bedding		Site pole position (from b)	
			D	*I*	*D*	*I*	Lat. (°N)	Long. (°E)
1	Kyloe Hills, Belford, natural cliff outcrop	NU 042 395	177	+ 2	176	−10	39°	183°
2	Belford, natural cliff near Craggyhall	NU 103 346	185	− 2	185	− 6	37°	172°
3	Belford, quarry	NU 110 346	195	− 5	196	− 5	35°	158°
4	Belford Station quarry	NU 129 342	178	−11	179	− 4	37°	180°
5	Bamburgh, sea cliffs, near lighthouse	NU 173 360	187	− 7	186	− 5	37°	171°
6	Bamburgh, sea cliffs, under castle	NU 184 352	192	+ 7	192	+ 3	32°	164°
7	Budle quarry, near Bamburgh	NU 160 342	183	− 2	183	− 1	35°	175°
8	Embleton quarry	NU 232 230	203	−10	203	−11	36°	150°
9	Dustanburgh castle, sea cliffs	NU 260 221	191	−22	193	−28	48°	159°
10	Craster harbour, sea cliffs	NU 259 200	169	−16	171	−14	38°	190°
11	Craster quarry, sea cliffs	NU 257 197	190	−17	191	−14	40°	164°
12	Cullernose Point, sea cliffs	NU 260 187	189	−25	190	−20	44°	165°
13	Longhoughton, Howlet Hills quarry	NU 234 956	187	−15	187	−15	42°	169°
14	Longhoughton, Snableazes quarry	NU 224 139	199	−22	203	−26	46°	153°
15	Newton on the Moor, Hampeth burn quarry	NU 154 075	196	−19	196	−19	44°	170°
16	Rothbury, Ward's Hill quarry	NU 078 964	205	+ 6	205	+ 6	28°	150°
17	Rothbury, Coltpark quarry	NZ 061 941	182	+ 7	182	+ 9	30°	176°
18	Knowesgate quarry	NY 991 858	189	− 5	190	−13	41°	165°
19	Knowesgate, West Whelpington Crag's quarry	NY 977 838	183	−10	183	+ 1	34°	174°
20	Kirkwhelpington river cliffs	NY 999 843	193	+28	190	+15	37°	167°
21	Kirkwhelpington, Three Farms quarry	NY 988 834	184	+12	183	+ 7	31°	174°
22	Kirkwhelpington, Northside natural outcrops	NY 986 824	180	+15	179	+11	29°	179°
23	Great Bavington, natural cliff outcrop	NY 982 807	189	+14	190	+13	38°	167°
24	Little Bavington, Divethill quarry	NY 980 790	183	− 9	177	−10	40°	182°
25	Thockrington, natural cliff outcrop	NY 957 790	184	−17	195	+11	28°	161°

Table 1—*continued*

Site No.	Locality	National Grid reference	Site Directions of Magnetization (a) referred to present horizontal		(b) corrected to bedding		Site pole position (from b)	
			D	*I*	*D*	*I*	Lat. (°N)	Long. (°E)
26	Gunnerton quarry, near Barrasford	NY 912 744	182	+12	182	+11	29°	176°
27	Twice Brewed, natural cliff outcrop	NY 754 676	182	+ 7	182	+ 7	31°	175°
28	Greenhead, quarry near Milecastle 42	NY 715 667	192	+ 9	190	−15	42°	164°
29	Greenhead, Walltown quarry	NY 671 661	183	−15	183	−20	45°	173°
30	Stanhope, Greenfoot quarry	NY 982 391	194	−11	194	− 7	38°	160°
31	Stanhope, Tees river bed	NY 988 391	186	−10	186	− 6	38°	170°
32	Copthill, quarry (laccolith)	NY 836 419	202	+18	202	+18	—	—
33	High Force, Tees river cliff	NY 872 281	192	− 7	192	− 9	38°	162°
34	Langdon Beck quarry 1	NY 871 291	—	—	—	—	—	—
35	Langdon Beck quarry 2	NY 875 292	192	+ 3	192	0	35°	163°
36	Middleton in Teesdale, Crossthwaite quarry	NY 836 418	181	+11	180	0	35°	178°
	Mean values ($N = 34$)				187°·8	−4°·9	37°·3	168°·9

Table 2

Site	Maximum i.r.m. gauss ($\times 10^{-1}$)	Field required to saturate (oersteds)	Coercivity of remanence (oersteds)	n.r.m. gauss ($\times 10^{-4}$)	Susceptibility in $H = 0{\cdot}49$ oe e.m.u. ($\times 10^{-4}$)	Ratio of n.r.m. to susceptibility
7	6·9	1 000	270	19·9	22·5	0·9
10	13·1	1 000	350	31·0	21·6	1·4
12	18·9	1 800	390	40·0	23·1	1·7
15	10·0	1 300	330	28·6	16·4	1·7
24	7·5	800	270	20·2	15·6	1·3

in Table 2 are similar to those measured by Irving (1956) in the Tasmanian dolerite sills which are of similar composition. The same is true of the intensities of n.r.m.

5. Contact sediments and origin of magnetization of the sills

At three sites (16, 17 and 19) samples of sedimentary rocks near the lower Whin contact have been collected. The upper contacts when exposed are generally inaccessible. At site 19, a continuous section was obtained through the exposed thickness of two feet of fine-grained yellow-brown sandstone. At site 17, a greater

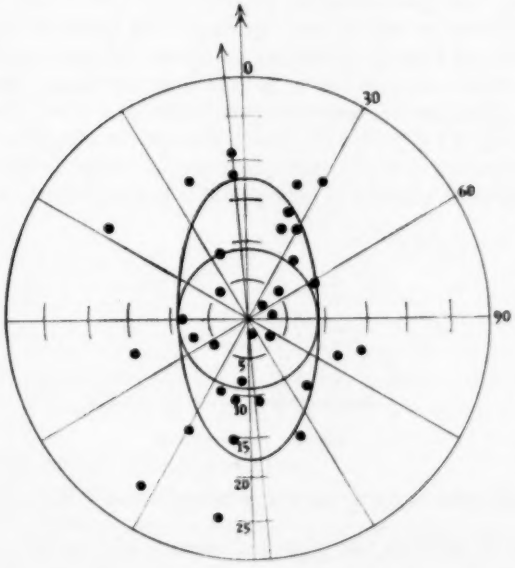

FIG. 3.—Centred stereographic plots.
Site mean directions of magnetization. The semi-axis of the oval (9°·7 and 19°·3) have been calculated from the standard deviation of the distribution of poles as described in paragraph 8. 21 points lie inside this oval and 13 outside.

thickness of more massive ganister-like sandstone is exposed and samples were taken at the contact and 18 in. lower down. Similarly at site 16 two samples of bluish limestone were collected at, and 18 in. below, the contact. The angular differences between the mean directions of the contact sediments and the Whin Sill at each of the three sites were 4, 14 and 16° respectively. Close to the contact the intensity of n.r.m. is about 10^{-5} gauss and 2 ft away is about 10^{-6} which is an order greater than in unheated sediments of the same type. The latter are so weakly magnetized as to be only just measurable. The higher n.r.m. parallel to that of the sills found in the contact sediments suggests that both magnetizations are thermoremanent in origin and that they consequently date from the time of injection and cooling of the latter.

6. Stability of n.r.m. of the Whin Sill

Ten specimens were randomly selected and the directions of n.r.m. were measured again after an interval of one year, having been stored in the Earth's field during that time. The differences between original and repeat directions of magnetization of the specimens were 1°, 13°, 4°, 15°, 15°, 7°, 26°, 7°, 4°, and 0°. Although these may seem unduly great, the difference of 7° between the original and repeat mean directions of this group of specimens is not significant as the standard errors are 6°·2 and 5°·3 respectively. These differences are probably due to a randomly directed temporary magnetization acquired during storage in the laboratory.

Ten different specimens from the same sites as the remeasured samples were subjected to a.c. demagnetization as described by Creer (1958 and 1959). Demagnetization curves for two of these specimens are shown in Figure 5 (curves 1 and 2) the other eight being cleaned by an alternating field of 280 oersteds peak value without treatment in fields of intermediate strength. Specimens showed differences in direction of magnetization before and after cleaning of 9°, 7°, 5°, 9°, 23°, 35°, 6°, 6°, 5° and 17°. These changes in direction are irregular so that the difference between the mean direction of magnetization of these specimens before and after cleaning is only 3° and is not significant.

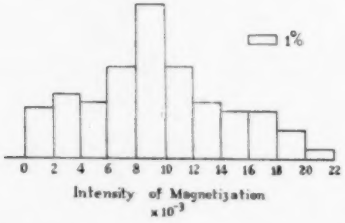

FIG. 4.—Histogram of the intensity of magnetization of Whin Sill samples.

The specimens showing the biggest changes in each of the above tests came from site 32, and because of doubt as regards the stability of this rock, no pole position has been calculated from the site mean direction of magnetization, nor has the latter been used in the statistical computation of the mean direction of magnetization of the whole formation.

In Figure 5, curve 3 illustrates the behaviour of a highly unstable rock (a Triassic basalt from Uruguay studied by Creer 1958), a marked decrease in intensity and a marked change in direction of magnetization being produced by demagnetizing fields of peak value less than 100 oersteds. The Whin Sill specimens did not behave in this way and the magnetization of this formation is therefore believed to be stable within the standard deviation of the distribution about the mean.

7. Length of record

In a single sill all parts do not cool through the Curie point at the same time. Jaeger (1957) has shown, on reasonable assumptions, that after injection the margins cool to about 665 °–620 °C which is somewhat higher than the Curie temperature of magnetite (580 °C). Thereafter the temperature falls slowly, the time taken

for the Curie point isotherm of magnetite to pass from the margin to the centre being approximately $0{\cdot}006d^2$ years where d is the thickness in metres. For the Whin Sills this is of the order of 10 years. Thus the average of direction from several specimens taken from one locality—say for instance a quarry—will approximate to a spot reading of the field direction. However it is unlikely that the sills were injected at the same time, and in some places there is evidence of multiple intrusion (Westoll & others 1955, Smythe 1914). The time spread is probably not less than a hundred years and is more likely to be several thousand years. Therefore the directions of magnetization at different sites are likely to be affected by the secular variation of the field.

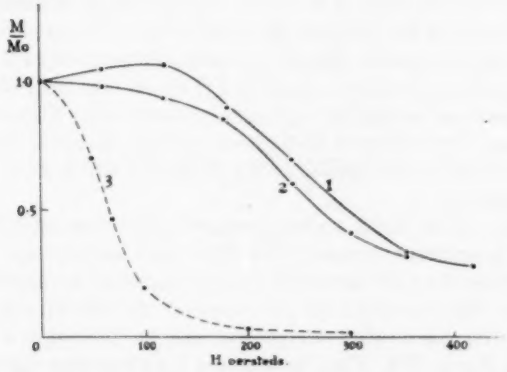

FIG. 5.—Demagnetization curves of Whin Sill (curves 1 and 2) and Uruguay basalt (curve 3) samples.

8. Analysis of directions of magnetization

It has been customary to use the distribution suggested by Fisher (1953) to compute the mean direction and confidence limits for palaeomagnetic directions obtained from a rock formation. From this it is an easy matter to calculate the attitude of the mean geocentric dipole (relative to the present geographical co-ordinates) in the field of which the rock formation became magnetized (Creer, Irving & Runcorn 1957). Fisher's distribution supposes that the individual directions, when regarded as points on a unit sphere, are distributed over the sphere with probability density

$$K \exp(K \cos \theta)/4\pi \sinh K \qquad (1)$$

where θ is the angle between the mean direction (direction cosines λ, μ, ν) and the direction of a specimen (l, m, n) and is given by

$$\cos \theta = \lambda l + \mu m + \nu n \qquad (2)$$

the individual directions are symmetrically disposed around the mean direction so that the contours containing given percentages of the observations are concentric circles with their centres at the true mean direction. Their radius θ_0 in radians is given by

$$\theta_0 = \{-2/K \,.\, \ln[\text{Prob}(\theta > \theta_0]\}^{\frac{1}{2}} \qquad (3)$$

provided θ_0 is fairly small.

An estimate k of K is given by

$$k = N-1/N-R \tag{4}$$

where

$$R^2 = (\Sigma l_i)^2+(\Sigma m_i)^2+(\Sigma n_i)^2. \tag{5}$$

For a Fisherian distribution the angle in radians equal to $(2/K)^{\frac{1}{2}}$ is analagous to the standard deviation σ in a Gaussian distribution and in the following will be termed standard deviation and denoted by θ_0. Thus

$$\theta_0 = 81/K^{\frac{1}{2}} \text{ degree} \tag{6}$$

as has previously been derived by Runcorn (1957) and, for N sufficiently large that $N \doteqdot N-1$, is the same as the analogue of standard deviation used by Wilson (1959). However, whilst for a Gaussian distribution about 68 per cent of the measurements lie within the standard deviation, equation (3) shows that only about 63 per cent of the measurements lie within the analogous quantity for a Fisherian distribution.

It can be seen, by inspection of Figures 2 and 3, that this distribution is not applicable to the directions observed in the Whin Sill which have an oval and not a circular distribution.

The directions of the Earth's palaeomagnetic field measured for a number of sites do not give a continuous record of the field but a series of spot readings of the field during the time the rock formation in question became magnetized. Thus, as an alternative we may represent the directions at one site by a geocentric dipole (f), and assume that this dipole undergoes secular oscillations in a random fashion around the mean dipole (F). Thus in equation (1) θ becomes the angle between f and F, and the distribution of observed directions of magnetization will be oval, the semi-axes depending on the palaeomagnetic latitude of the rock formation as follows:

$$\delta D = \theta_0(1+3\cos^2\psi)^{-\frac{1}{2}} \tag{7}$$

and

$$\delta I = 2\theta_0(1+3\cos^2\psi)^{-1} \tag{8}$$

where ψ is the angular distance from the rock site to F (palaeomagnetic colatitude) and δD and δI are the changes in palaeomagnetic declination and inclination caused by movement of the pole from F to f. In low palaeomagnetic latitudes, δD tends to $81K^{-\frac{1}{2}}$ and δI tends to $162K^{-\frac{1}{2}}$ for $P(\theta < \theta_0) = 0{\cdot}63$. In high geomagnetic latitudes both δD and δI tend to $40{\cdot}5\ K^{-\frac{1}{2}}$. Thus, given a circular distribution of poles (f), the observed directions of magnetization of rocks from low palaeomagnetic latitudes should have oval distributions with $\delta D = \frac{1}{2}\delta I$ while in rocks from high palaeomagnetic latitudes the observed directions of magnetization should have a circular distribution, the variation in inclination and declination being one-half the variation for declination in low latitudes.

The inclination of the mean direction of magnetization in the Tasmanian dolerites is 85° indicating formation at high latitudes, and Watson & Irving (1956) found that the site directions of magnetization were distributed about the mean in a Fisherian fashion. The results presented in this paper for the Whin Sill provide data for low palaeomagnetic latitudes. The distribution of site directions of magnetization can be seen to be oval (Figure 3) and not circular. However, poles (f) for each of the 34 sites have been calculated and are plotted on a

stereogram with the mean pole F at the centre of the projection. The distribution is seen to be more nearly circular (Figure 6) which suggests that it is more correct to apply Fisherian statistical methods to site pole positions and not to site directions of magnetization. This has been done and the results are given in Table 3. In Figure 6 a circle of radius 9°·1 equal to the standard deviation for the pole distribution is drawn and it can be seen that this contains the expected proportion

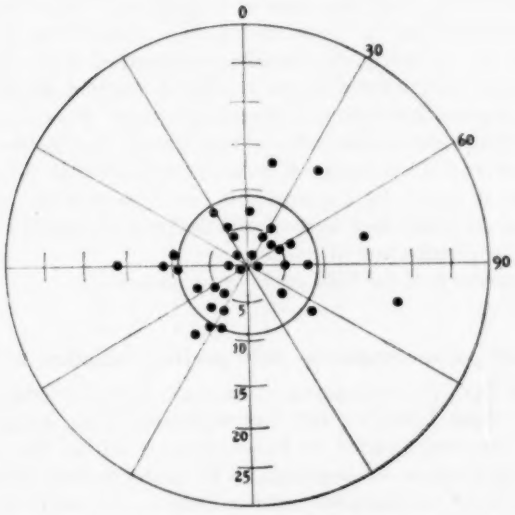

FIG. 6.—Site pole positions with the standard deviation circle (radius 9°·1). This circle contains 23 out of the 34 points.

(63 per cent) of the site poles. From equations (7) and (8) the values of the semi-axes of the standard deviation of the poles may be obtained. These are 9°·1 and 18°·2 respectively. This oval is drawn in Figure 3, and again it contains approximately 63 per cent of the observations. The pole calculated from the mean directions of magnetization of sites (168° E 37° N) does not differ appreciably from the mean pole position given in Table 3 since the mean of an oval distribution is still the sum of the direction cosines as for a circular distribution.

Table 3

The mean pole position of the Whin Sill

Latitude	Longitude	N	K	Standard deviation θ_0	Standard error θ_0	Error $P = 0{\cdot}05$ α
37°·3 N	168°·9 E	34	79·5	9°·1	1°·6	2°·8

There are several possible reasons why the distribution of directions in Figure 3 is oval.

(a) It was suggested by Graham & others (1957) that due to magnetostriction a n.r.m., acquired under non-hydrostatic load (e.g. in certain igneous rocks such as

sills) changed on release of the stress due to erosional unloading or removal of the sample from the outcrop, and it is conceivable that such effects could have caused the oval distribution.

(b) The oval distribution may be due to a superimposed secondary component of the type discussed by Creer (1957). The most likely direction for such a secondary component is that of a dipole field along the present axis of rotation in which case the long axis of the oval distribution should be along the great circle connecting this direction to the mean direction of magnetization calculated above for the Whin Sill. This great circle is shown in Figure 6. Unfortunately it makes only a small angle with the palaeomagnetic meridian also marked in this figure so that it is not possible to say on geometrical grounds alone whether the data agree better with this interpretation than with that considered above. However the close agreement between many pole determinations from Europe (see Section 9) using rocks of different types and stress histories makes it unlikely that either (a) or (b) are important. Also it would be a remarkable coincidence if the effects of partial instability and stress deflections were of just the right magnitude and direction to produce a circular distribution of poles.

(c) Secular variation of the field could be responsible.

9. The Permian palaeomagnetic pole position relative to Europe

Data derived from palaeomagnetic studies of several formations in the Lower Permian of the British Isles, Norway, Germany and France are given in Table 4. The period of time represented probably spans much of the Lower Permian (Autunian and Saxonian = Rotliegende). The dating (which is by normal stratigraphical methods) of the intrusive igneous rocks is necessarily a little doubtful, especially with regard to some of the Norwegian and Ayrshire rocks. In the case of the Esterel igneous rocks and the Whin Sill, it is possible that some portion of the uppermost Carboniferous (Stephanian) is represented. The length of time covered by all formations is probably of the order of 10 million years. The calculated pole positions all lie in the Northern Pacific (Figure 7b). The site poles for the Whin Sill are given in Figure 7a for comparison and it is seen that they have a similar distribution.

It is of interest to investigate whether the apparent differences between these pole positions (Figure 7a) are significant or not. Such an investigation would give information about post-Permian relative movements between sampling areas or the extent of polar movements during the time represented. To do this it has first to be asked whether directions of magnetization have been sampled from a time interval large enough such that the means may be regarded as main dipole field directions. For those areas for which adequate data is available the standard deviation of the distribution of site mean directions of magnetization has been calculated in the same way as described for the Whin Sill in paragraph 4, and this is taken to represent the effects of secular variation during the interval of time when the formation in question acquired its magnetization. Consequently, it may be inferred that data for which the standard deviation of the distribution of site means is too small does not cover the full range of the ancient secular variation and therefore any geomagnetic pole calculated from it may be of the order of 10° away from the palaeogeographical pole. Reasons are given in paragraph 8 for suggesting that the length of the record revealed by the present work on the Whin Sill probably covers the full range of secular variation and the standard deviation of the

Table 4

	Locality			Formation	Directions of n.r.m.		Standard deviation	Standard error	Number of samples	Number of sites	Position of Palaeomagnetic pole		Reference
					D	I	θ_0	θ_e	N_r	N_m			
British Isles	Ayrshire	55°·5 N	4°·5 W	Minor (Kylitic) intrusions	181°	+ 7°	16°·6	7°·4	11	6	32° N	174° E	Armstrong 1957
	Ayrshire	55°·5 N	4°·5 W	Mauchline sandstones	187°	− 6°	?	4°·6*	26	?	—	—	Du Bois 1957
	Ayrshire	55°·5 N	4°·5 W	Mauchline lavas	180°	− 4°	?	6°·7*	34	?	—	—	Du Bois 1957
	Northumberland and Durham	55° N	2° W	Whin Sill	188°	− 5°	13°·9	2°·4	102	34	37° N	169° E	This paper
	Devonshire	51° N	4° W	Exeter volcanic traps	189°	− 9°	20°·9	10°·4	33	5	43° N	164° E	Creer 1955, 1957
Norway	Oslo	60° N	10° E	Lavas & rhomboporphyry	202°	−33°	?	5°·5*	12	?	—	—	Rutten & others 1957
Germeny	Pfalz	49° N	7°·5 E	Rotliegendes Ro	190°	+ 6°	16°·9	6°·0	13	9	38° N	175° E	Nairn (unpublished)
	Baden Baden	49° N	8° E	—(sandstones Ro-Rm)	184°	− 7°	12°·6	5°·5	10	6	45° N	182° E	Nairn (unpublished)
	Saar	49° N	10°·5 E	—R	181°	+ 4°	23°·2	11°·6	9	5	40° N	188° E	Nairn (unpublished)
	Freiburg	48° N	8° E	and lavas)—R	160°	+12°	18°·6	18°·6	4	2	—	—	Nairn (unpublished)
France	Vosges	48° N	6° E	Saxonian rhyolites	194°	− 6°	17°·7	7°·9	14	4	45° N	167° E	Nairn 1947
	Morvan	47° N	4° E	Saxonian sandstones					3	2			
	Esterel	43°·5 N	7° E	Pyromeride R_4	210°	−16°	?		5	?	—	—	Roche 1957
	Esterel	43°·5 N	7° E	Dolerites	175°	−13°	?		3	?	—	—	Roche 1957
	Esterel	43°·5 N	7° E	Rhyolite R_3	217°	+23°	?	4°·5	14	?	—	—	Rutten & others 1957

* Computed from 95 per cent error and given in the references.

distribution of site mean directions of magnetization for this formation is 13°·9.* Creer (1955) has estimated the scatter to be expected due to present-day magnitude secular variation effects for different latitudes on the assumption that the distortions in the present field from that of an axial dipole field around any line of latitude are the same that any point on that line of latitude would experience over a time long with respect to the time constants of the core motions responsible for the secular variation. He obtained a value of $\theta_0 \simeq 20°$ for equatorial latitudes. This is comparable with many of the values* shown in Table 4, though some are rather lower. Whether this is because secular variation effects were less than at present during the time of magnetization of these rocks or because the time interval represented by them is short is unanswerable. Clearly at the present

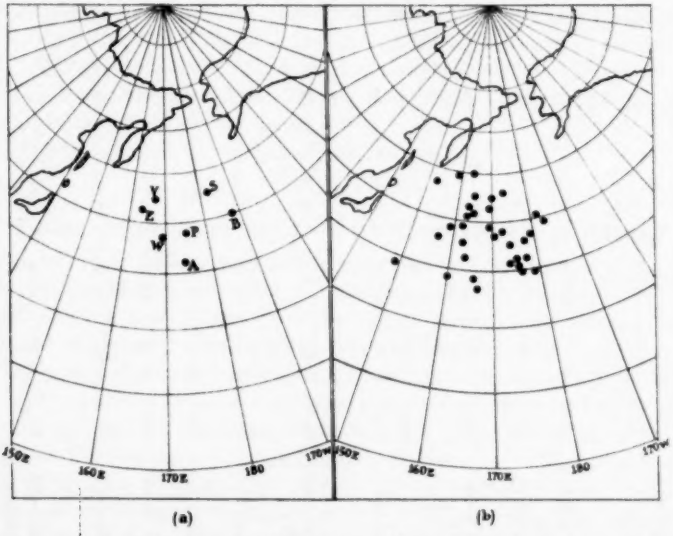

FIG. 7.—Maps showing the distribution of Lower Permian pole positions (a) and Whin Sill site pole positions (b).

stage it is not possible to say with certainty whether the differences between the pole determinations plotted in Figure 7a are real or are due to sampling inadequacies and although standard errors have been computed where possible, these should be treated with reserve at the moment, especially where the standard deviation is small or not given. The standard error used here is related to the standard deviation in the usual way and by the following formula $\theta_e = \theta_0/N_m^{\frac{1}{2}}$ where θ_e is the standard error, θ_0 the standard deviation of the distribution and N_m the number of sites which have been used in the computation of θ_0. The standard deviation is a measure of the width of the distribution and is one of its characteristic properties. With observations from a greater number of sites it can be determined with greater precision, but is not reduced numerically. (N_m must of course be large enough to represent the full range of secular variation as described above.) The standard error is a measure of the amount by which the

* In Table 4 θ_0 and θ_e have been calculated for the distribution of directions of magnetization and not from the distribution of poles given by these directions. This is because sufficient data was not available for recalculating all the localities quoted. For this reason it should be noted that the value given for θ_0 for Whin Sill differs from that given in Table 3.

calculated mean direction may differ from the true direction and is reduced as the number of sites is increased. For a Fisherian distribution it is the 63 per cent error cone. 95 per cent errors (as often quoted in palaeomagnetic work) can be computed from the data for θ_0 given in Table 4 using the following formula adapted from Irving & Watson (1957):

$$\alpha = 140/(KN_m)^{\frac{1}{2}} \tag{9}$$

and also equation (6).

10. Comparison with palaeoclimatic evidence

In Figure 8, Europe is drawn in the latitude and with the orientation suggested by the mean Permian result. On it are plotted the various climatic criteria; evaporites, desert sandstones, reef limestones, red beds and coal. These deposits

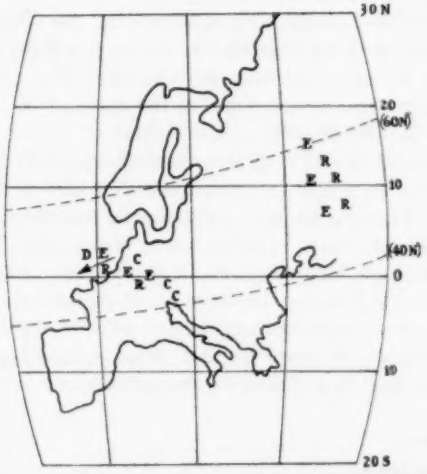

FIG. 8.—Map of Europe restored to the latitude and orientation suggested by Lower Permian palaeomagnetic results. The present 40° N and 60° N latitudes are retained for reference. Palaeoclimatic data are plotted using the following abbreviations: R reef corals, D desert sandstones, E evaporite and C coal deposits, arrow indicates the wind direction where known.

are usually thought to have been laid down in hot climates and their position close to the palaeomagnetic equator is consistent with the view that the geographic and mean magnetic poles coincide. (See general discussion in Irving 1956.) The wind direction inferred from the cross bedding in various dune sandstones (see Shotton 1956, Opdyke & Runcorn 1959) has a similar relation to the palaeomagnetic equator as the present day northern trade winds have to the present geographic equator. This is consistent with the proposed latitudes (see also Schove & others 1959) and suggests that the pole placed in the present day Pacific Ocean was the past geographic north pole assuming the present sense of rotation.

11. Conclusions

The Great Whin Sill has a stable, permanent magnetization with the site mean declination being 187°·8 with an inclination of −4°·9 confirming the earlier

work of Manley (1949) and Hallimond & Butler (1949). The mean pole position in longitude 168°·9 E and latitude 37°·3 N is consistent with the recorded Lower Permian pole positions in western Europe.

The method of analysis used in this paper always gives circular polar errors, and has the advantage that correction is made for differences in latitude and longitude of sites. This is particularly useful when results from formations with great lateral spread are analysed, or when comparisons are made between the results from formations laid down during the same geological period in different parts of a continental land-mass.

The model proposed is broadly consistent with current models of the Earth's field, namely that of a steady axial dipole component constant for periods of 10^6 years perturbed by non-dipole components with periodicities of the order of 10^2–10^3 years. The strength of the dipole field at the poles is twice that in low latitudes whereas the non-dipole components are irregular and their magnitude does not appear to depend on latitude. Thus, in general, the effect of the latter on total direction of the field may be expected to be less in high latitudes than in low latitudes, as would be expected from (7) and (8). The oval distribution in the Whin Sill may therefore owe its form to the manner in which the non-dipole components offset the dipole field in low latitudes.

Oval distributions of n.r.m. may have been obscured in previous work because of random errors due to relative rotations between sites or inherent dispersion of magnetization or partial instability, and because sampling has not usually been undertaken in sufficient detail. Furthermore, where thick sequences have been studied, covering perhaps many millions of years, polar wandering effects (movements of the dipole axis (F)) may be superposed on the non-dipole variations. In the Whin Sill the time range covered is probably of the order of hundreds or thousands, not millions, of years so that polar wandering is unlikely to have occurred during the time it acquired its magnetization.

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Gravity Anomalies and Crustal Structure in South Africa

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Summary

In this paper an account is given of the isostatic anomaly field in South Africa. Estimates of crustal thickness are made by using the regression coefficient criterion proposed by Heiskanen, the minimum sum of squares criterion proposed by Bowie and a new criterion based on minimum sum of squares of departure from the regression lines. It is found that the regression coefficient criterion is not satisfactory because it is sensitive to systematic effects such as arise in South Africa near the escarpment. The other criteria yield results which are consistent with the seismically inferred crustal structure, i.e. with a crustal thickness 30–40 km.

1. Introduction

In 1958 the Geological Survey of the Union of South Africa completed a regional gravity survey of the Union, and the results were published in the form of Bouguer anomaly contours overprinted on the $1:10^6$ geological map of South Africa. A complete account of the work including a list of the observations is to be published as a Geological Survey memoir. This regional survey consisted of some 6 300 gravity meter observations superimposed on the primary network of forty pendulum stations observed within the Union by the writers in 1948–49 (Hales & Gough 1950). The observations made during the regional survey have been reduced isostatically. The reductions were made on the basis of five distributions of the isostatic compensation. These are the Airy–Heiskanen hypothesis with sea-level crustal thicknesses of 20, 30, 40 and 60 km, and the Pratt–Hayford hypothesis with depth of compensation 114 km. The method of isostatic reduction used for the gravity meter observations was that described in the account of the pendulum observations (Hales & Gough 1950). Briefly, the procedure was to calculate the effect of the compensation in Hayford zones A to O and of the topography and compensation in zones 18 to 1, at all points of a grid of side one-half degree covering the Union. The isotatic corrections for the individual gravity meter stations were then found by linear interpolation within each half-degree square.

It is convenient to remark here that gravity anomalies, whether they be Bouguer or isostatic, indicate departures of the actual gravitational field from that predicted by a particular model of crustal structure. Bouguer anomalies, for instance, represent the difference between the actual gravity field and that corresponding to an Earth made up of ellipsoidal shells within each of which the density is

constant. In the Airy–Heiskanen type of isostatic hypothesis it is assumed that the crust has density 2·67 g/cm³ and the material under it density 3·27 g/cm³ and that surface inequalities, however small, are hydrostatically supported by appropriate variations in the thickness of the crust immediately under them. Again, the Pratt–Hayford type of hypothesis supposes that all columns of unit area from the topographic surface down to a surface 114 km below this surface are of equal mass, the density being uniform within a column and least where the topography is highest.

Airy–Heiskanen isostatic anomalies may arise in several ways. In the first place, the density of some part of the crust may differ from the value of 2·67 g/cm³ assumed in the calculations. Secondly, it is possible that the assumption of a single-layer crust may be an oversimplification. Recent seismic work in France, Germany, the United States and South Africa has tended to support the view that the continental crust is composed of two layers. In this case there would be no reason to believe that the interface between the granitic and intermediate layers would be less irregular than the Mohorovičić interface itself. In the third place, the crustal thickness may not vary strictly in accordance with the assumption of local compensation.

It is not surprising that there are departures from isostatic equilibrium, for these are the natural consequence of the finite strength of the crust of the Earth (Hales 1955). In general, the existence of an isostatic anomaly implies the existence of stresses in the crust or mantle to support the excess or deficiency of load. Investigations in various parts of the world have, however, shown that sufficiently large areas of the crust are, on the whole, isostatically compensated (Heiskanen & Meinesz 1957, Chapter 7).

In our view there is a considerable advantage in using isostatic anomalies when considering large scale structure. Regional effects due to changes of structure associated with topography largely disappear when isostatic anomalies are used. This is particularly important near steep coast-lines, for the gravity effect of the thinning of the crust at the continental boundary extends to considerable distances inland. To put the matter in another way, we regard the use of isostatic anomalies as the best method of regional correction for surveys of continental extent.

2. Crustal thickness: Heiskanen's criterion

Two methods have been proposed for estimating crustal thickness from gravity anomalies. Heiskanen (1924) has suggested that the most probable value of the crustal thickness is that for which the anomalies show the least correlation with height, or alternatively is the thickness for which the regression coefficient of anomaly against height is zero. On the results of the pendulum survey the following values of the correlation between anomaly and height were found (Hales & Gough 1950):

Table 1

Correlation between anomaly and height

Hypothesis	Free Air	AH20	AH30	AH40	AH60	H114	Bouguer
Correlation Coefficient	+0·275	+0·049	−0·002	−0·026	−0·063	−0·002	−0·900
P	0·05	0·73	0·99	0·85	0·66	0·99	0·00

In this table P is the probability that a correlation coefficient greater than that quoted would arise through random variation in uncorrelated data. It was remarked (Hales & Gough 1950) that "the correlation coefficients are significantly different from zero only for the Free Air and Bouguer hypotheses which are equivalent to compensation at zero and infinite depths respectively", and the conclusion was drawn that the topography was compensated at some depth between zero and infinity.

In 1954, Hales & Gough (1954) made an analysis of the gravity meter observations at a stage when the regional survey was incomplete. From the two to three thousand observations then available the following regression coefficients of anomaly on height were found:

Table 2

Regression coefficients of anomaly against height in mgal per km

Hypothesis	Free Air	AH20	AH30	AH40	AH60	H114	Bouguer
Regression Coefficient	+19	+9	+6	+5	+1	+3	−90
99 per cent Confidence limits	±7	±5	±5	±5	±5	±5	±7

The implication that the sea-level crust had thickness greater than 60 km was in conflict with the results of seismic studies of crustal thickness in the Transvaal. It was noted, however, that there was a quite marked tendency for negative anomalies to occur at the foot of the great escarpment. In addition, it so happens that numerous positive anomalies due to geological structures occur on the high interior plateau. Effects of these kinds must introduce a positive correlation between anomaly and height, and will displace zero regression coefficients to a crustal thickness greater than the true thickness.

Since 1954 the Geological Survey has made many additional observations, especially in those regions which were only poorly covered by the observations made before that date. Figure 1 is a simplified version of a map of isostatic anomalies on the AH30 hypothesis, based on all the observations. In this map isostatic anomaly contours at 20 mgal intervals are superimposed on a topographic map with contour interval 500 m. It is clear that there is a general tendency for negative anomalies to occur below the escarpment, with a belt of positive anomalies at the top. In some regions the negative anomalies are very large over considerable areas. In the north-eastern Transvaal, for instance, the −40 mgal contour encloses an area roughly 240 km by 80 km and the anomaly fell to −60 mgal in places. Another region of marked negative anomaly is that in the south-eastern Cape, where again the −40 mgal contour encloses a large area in the form of a strip 240 km long and 50 km wide. In this case the anomaly falls to −80 mgal. In the eastern part of this anomaly part of the mass deficiency may result from a basin filled with Cretaceous sediments. Elsewhere in the southern Cape, however, the surface rocks are the compact quartzites, sandstones and shales of the Cape system, with densities not far from 2·67 g/cm^3. There is no doubt that this anomaly results not from near-surface geology but from a departure at depth from the simple Airy–Heiskanen crustal model for which the corrections were calculated. The

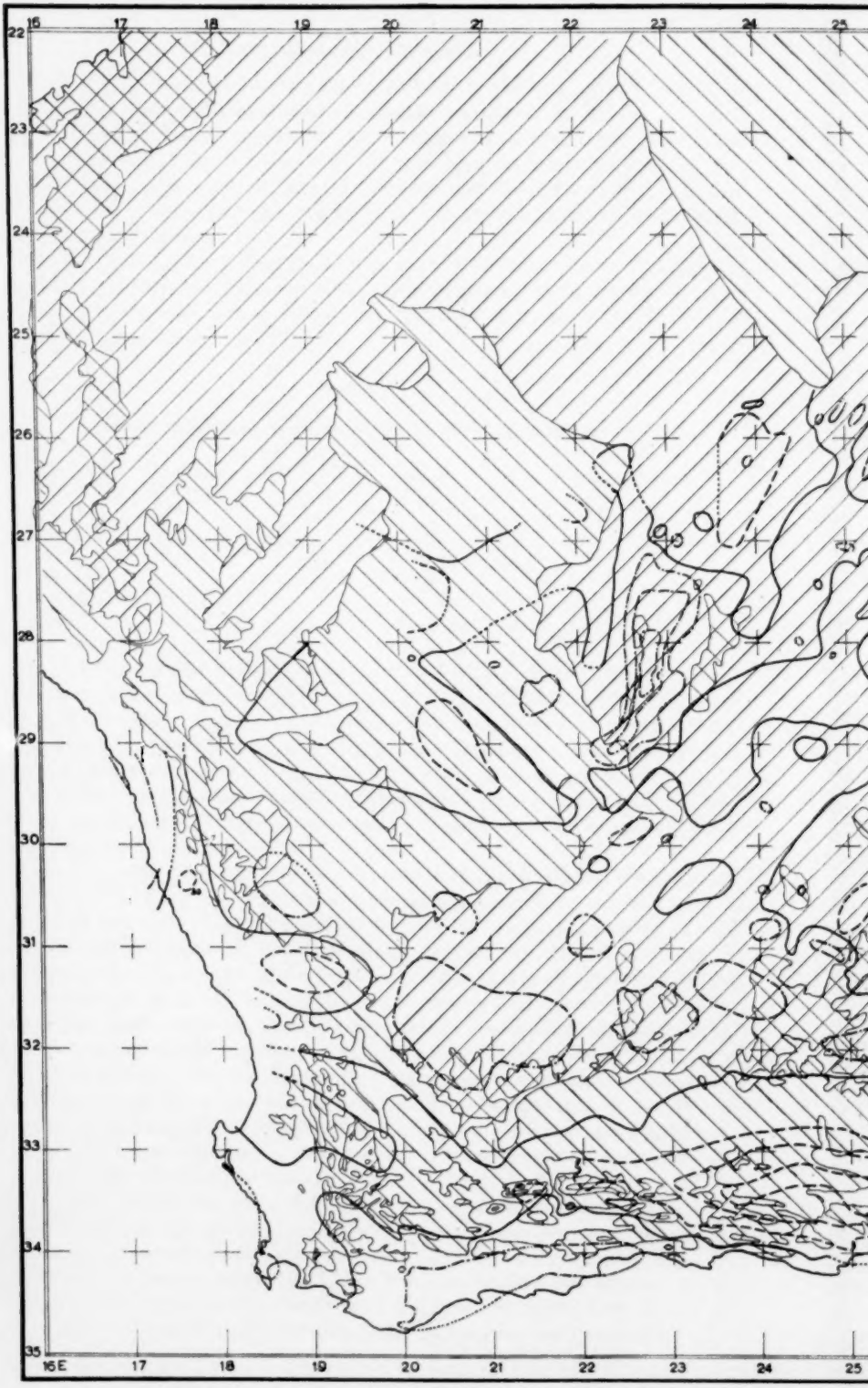
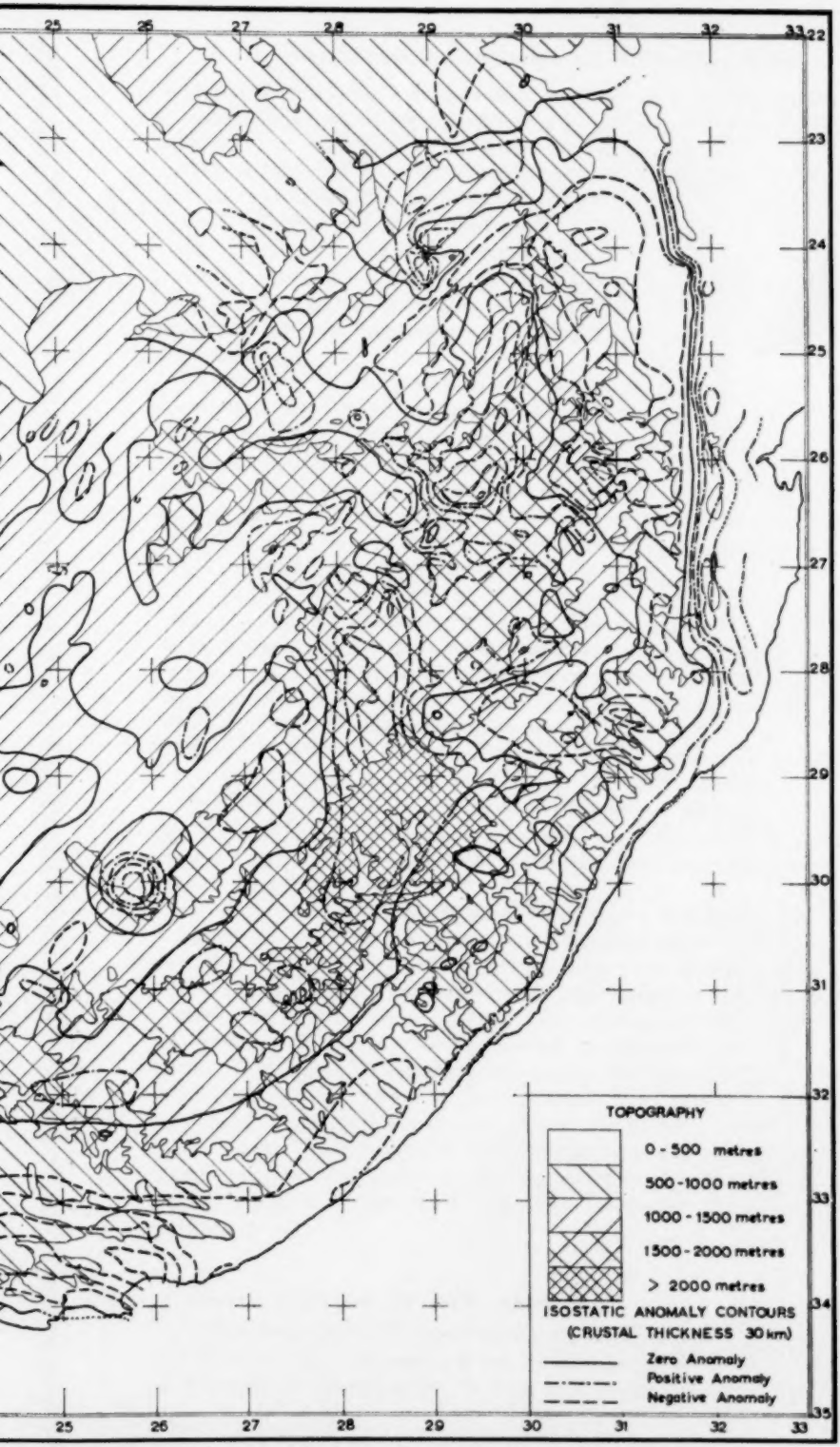


FIG. 1.—Isostatic Anomaly Map of the Union of South Africa. The isostatic anoma
intervals. The isostatic anomalies were calculate



ic anomaly contours are at 20 mgal intervals; the topographic contours at 500 m
calculated for a crustal thickness of 30 km.

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absence of steep gradients is consistent with this view. In the eastern Transvaal the negative anomalies occur in a granite–gneiss region and again it is impossible to account for the anomaly in terms of near-surface geology.

Having regard to the relationship between the anomalies and the escarpment, we think it probable that the negative anomaly is due to the compensation of the former topography which has been eroded away as the escarpment has retreated. If this view is correct it is clear that some of these negative anomaly areas have existed for times of the order of tens of millions of years. In a separate paper we shall examine this interpretation of the isostatic anomalies associated with the escarpment, and the implications with regard to the strengths of the crust and of the mantle.

In the study of the pendulum observations it had been found impossible to discriminate between the isostatic hypotheses with only 53 observations available. At the time the 1954 report was written there were between two and three thousand observations. It was recognized that the use of each observation with unit weight was not justified on account of the non-uniform distribution of the observations in relation both to area and to height. Consequently the unit of observation was chosen to be the "square" bounded by two latitudes one-quarter degree apart and two longitudes one-quarter degree apart. Every such quarter-degree square containing one or more observations was given unit weight, the mean anomaly and mean height for the observations in the square being used. In the final analysis based on 6 300 observations the same method has been used. To give a check on the possible effect of correlation between neighbouring quarter-degree squares, regression coefficients have also been calculated for all squares of side one degree within which there are observations.

Table 3 gives the regression coefficients and their mean square errors for 1 372 quarter-degree squares, and for 142 one-degree squares.

It will be seen that the regression coefficients are all positive for the isostatic hypotheses for the $\frac{1}{4}$ deg squares, and only the AH60 coefficient is negative for the 1 deg squares. The Heiskanen criterion therefore again indicates a crust about 60 km thick.

In order to test the hypothesis advanced above, that the regression criterion is rendered unreliable by the escarpment anomalies and by the various large geological anomalies, regression coefficients have been calculated for $\frac{1}{4}$ deg squares within strips lying between latitudes one degree apart. The results are given in Table 3 and represented in Figure 2. The number next to each set of regression coefficients in Figure 2 gives the latitude bounding the one-degree strip on the north. It will be seen that the sub-classes of the data, selected arbitrarily in this way, show wide variations of the regression. Zero regression occurs for almost any given crustal thickness in some part of the country. Furthermore, it will be noted that there are significant differences between the coefficients for the quarter and one degree square data. It appears, therefore, that this method of estimating crustal thickness is too sensitive to other systematic effects to be regarded as satisfactory.

3. Crustal thickness: Bowie's criterion

The second method for determining the most probable depth of compensation (on the Pratt–Hayford model), or crustal thickness (on the Airy–Heiskanen model), is that used by Bowie (1917). It rests on the assumption that the most probable

Table 3

Regression coefficients

Data	No. of observations	Regression coefficients in mgal/kiloft						
		Free Air	AH20	AH30	AH40	AH60	H114	Bouguer
Quarter-degree squares: Union of South Africa	1372	+30·9 ± 1·6	+12·3 ± 1·3	+ 9·7 ± 1·3	+ 7·5 ± 1·3	+ 3·6 ± 1·3	+ 6·5 ± 1·3	− 81·9 ± 1·6
One-degree squares: Union of South Africa	142	+20·0 ± 3·6	+ 8·1 ± 3·0	+ 4·6 ± 3·0	+ 3·0 ± 3·0	− 0·7 ± 3·3	+ 1·8 ± 3·3	− 93·7 ± 3·6
Quarter-degree squares:								
22–23° S	38	−13·2 ±16·4	−27·5 ±11·5	−32·3 ±11·2	−34·3 ±11·2	−40·5 ±10·8	−38·4 ±11·5	−108·9 ± 16·4
23–24° S	60	+45·8 ±12·5	+17·4 ±10·8	+10·7 ±10·8	+ 5·7 ±10·8	− 2·9 ±10·5	+ 8·1 ±10·8	− 63·8 ± 12·1
24–25° S	79	+46·2 ±10·2	+20·5 ± 9·2	+15·1 ± 8·9	+11·9 ± 8·9	+ 5·4 ± 8·9	+ 8·7 ± 8·9	− 64·0 ± 10·2
25–26° S	112	+46·4 ± 7·5	+13·2 ± 6·2	+ 9·3 ± 6·2	+ 2·2 ± 6·6	− 5·1 ± 6·6	− 0·5 ± 6·6	− 66·3 ± 7·5
26–27° S	142	+34·8 ± 5·6	+ 9·7 ± 5·2	+ 4·9 ± 5·2	+ 0·9 ± 5·2	− 7·3 ± 4·9	− 3·6 ± 4·9	− 81·7 ± 4·9
27–28° S	160	+10·6 ± 4·3	− 6·4 ± 3·9	− 9·1 ± 3·6	−12·2 ± 3·6	−17·5 ± 3·6	−14·8 ± 3·6	−101·9 ± 4·3
28–29° S	164	+13·4 ± 5·2	+ 1·4 ± 4·6	− 1·2 ± 4·3	− 3·7 ± 4·3	− 8·4 ± 4·6	− 6·7 ± 4·6	− 99·1 ± 5·2
29–30° S	127	+10·2 ± 4·3	− 6·5 ± 3·3	−10·8 ± 3·3	−14·7 ± 3·3	−21·2 ± 3·3	−17·4 ± 3·3	−102·7 ± 4·3
30–31° S	115	+12·5 ± 5·6	+ 6·2 ± 3·6	+ 1·8 ± 3·6	− 1·6 ± 3·6	− 9·1 ± 3·9	− 6·8 ± 3·9	−101·0 ± 5·6
31–32° S	112	+44·5 ± 4·3	+20·1 ± 3·3	+16·0 ± 3·3	+11·7 ± 3·3	+ 5·5 ± 3·6	+ 9·2 ± 3·6	− 68·8 ± 4·3
32–33° S	113	+53·5 ± 5·2	+21·8 ± 3·3	+18·7 ± 3·3	+16·0 ± 3·6	+11·3 ± 3·6	+16·2 ± 3·6	− 59·2 ± 5·2
33–34° S	113	+27·4 ± 9·8	− 5·2 ± 9·5	− 7·9 ± 9·5	− 8·8 ± 9·5	−11·8 ± 9·2	− 9·0 ± 9·5	− 84·4 ± 9·8
34–35° S	37	+56·1 ±20·7	+20·0 ±27·9	+18·7 ±31·2	+15·0 ±33·8	+13·8 ±37·7	+19·5 ±35·8	− 58·9 ± 20·7

The limits are standard errors.

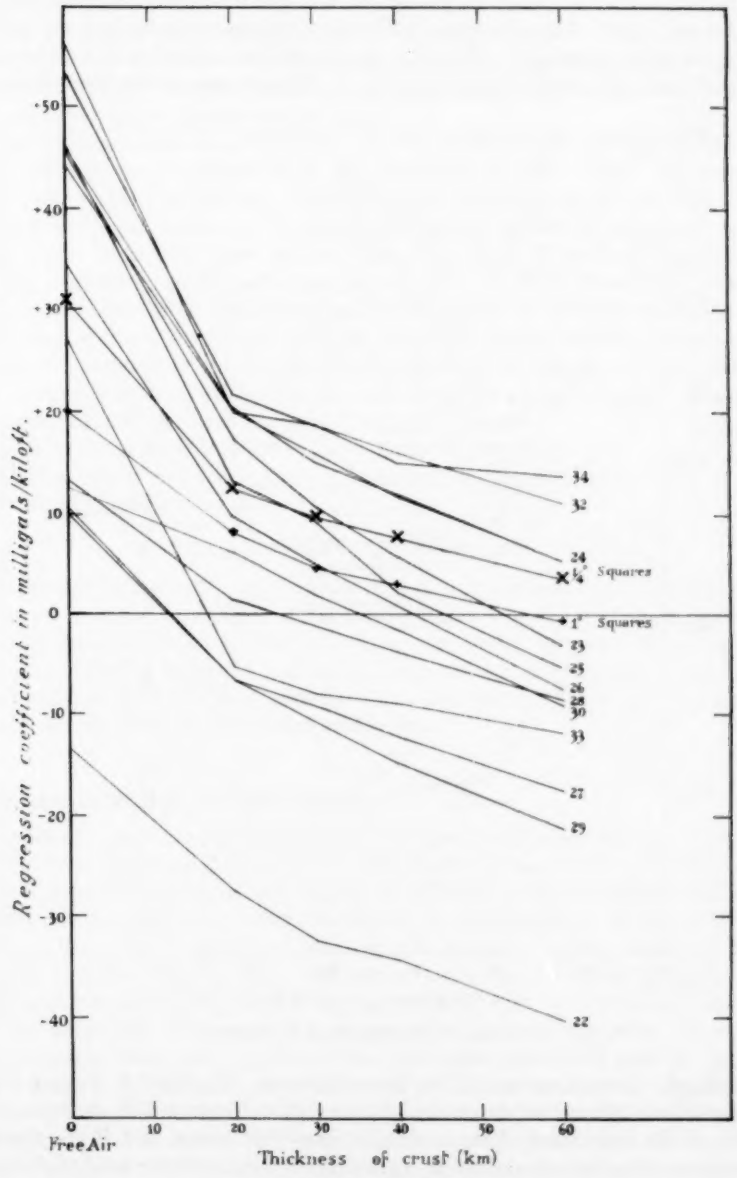

FIG. 2—Regression coefficients.

hypothesis is that for which the sum of the squares of the departures from the mean anomaly is least, that is, for which the variance is least.

The variances have been calculated for the quarter-degree means and for the one-degree means. They are given in Table 4. Figure 3 shows how the variance changes with the hypothesis. It will be seen that both curves show a minimum in the neighbourhood of 30 to 40 km thickness. The abscissa of the minimum is not

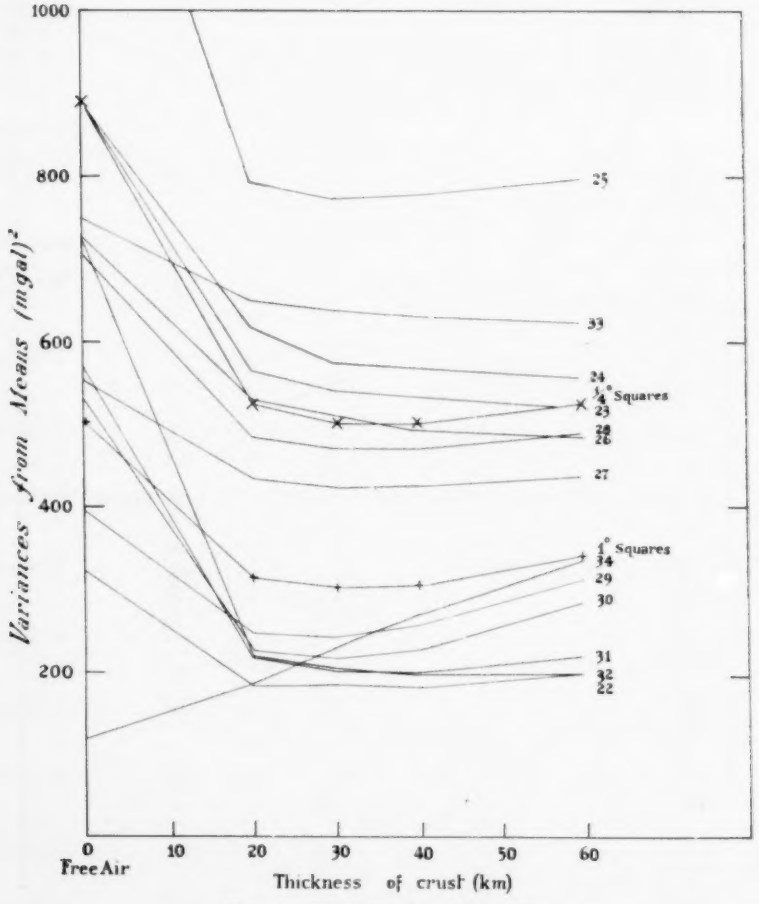

FIG. 3.—Variances from the mean.

well defined. In making use of the Bowie criterion, Horsfield & Bullard (1937) remarked that: "It can be shown that if the relation between $\Sigma\epsilon^2$, the sum of the squares of the departures of the anomalies from their mean, and H the depth of the compensation, be represented by a parabola in the neighbourhood of the minimum, then the standard error of the depth of compensation determined from the position of the minimum is $[\Sigma\epsilon^2/a(N-1)]^{\frac{1}{2}}$ where N is the number of stations and a is the coefficient of the term in H^2."

The parabola fitted to variances of the anomalies of the quarter-degree squares on the hypotheses, Free Air, AH20, AH30, AH40, and AH60, has a minimum at

41·8 ± 1·2 km; that fitted to the AH20, AH30, AH40 and AH60 variances has a minimum at 39·8 ± 2·4 km, and that fitted to AH20, AH30 and AH40 variances a minimum at 36·6 ± 2·0 km. The errors quoted are standard errors. The corresponding figures for the one-degree squares are 38·9 ± 4·0 km, 34·6 ± 5·8 km and 33·7 ± 5·3 km. It is noteworthy that in this case there is good agreement between the quarter-degree and one-degree curves with the exception that the variance of the quarter-degree square data is greater.

The variances have been calculated for the quarter-degree means in the one-degree latitude strips discussed in the preceding section. They are given in Table 4 and plotted in Figure 3. It will be observed that in a number of cases the curves show broad minima in the neighbourhood of 40 km. These strips are, as a rule, those in which there are no large anomalies. For these the Bouguer anomaly regression coefficient is in the neighbourhood of the usual value of −0·0031 mgal per foot. In other cases the minimum is not defined, but except in the cases of the strips for 23°, 24°, 26° and 33° S, the curves are consistent with a minimum variance in the vicinity of 30 to 40 km. It appears that the effect of large anomalies is, in general, to lower the resolution of the criterion. However, the variance from the mean consists of two components:

(1) The variance of the observations from the regression line.

(2) The variance of the regression line about the mean.

In fact it can be shown that

$$\Sigma(y-\bar{y})^2 = \Sigma(y-y')^2 + b^2\Sigma(h-\bar{h})^2.$$

In this equation $\Sigma(y-\bar{y})^2$, $\Sigma(h-\bar{h})^2$ are the sums of the squares of the departures of y and h from their means, y' is the value predicted by the regression line at the height of the observation y, $\Sigma(y-y')^2$ is the sum of the squares of the departures of the observations from the regression line, and b is the regression coefficient. It follows that the variance about the mean may be affected in some degree by systematic effects in the regression coefficients.

4. Crustal thickness: a new criterion

In fitting a curve to observations by least squares, the best values of the parameters of the curve are determined by the condition that the sum of the squares of the departures from the line shall be a minimum. This suggests that in the present case, in which one wishes to choose between different values of a parameter, namely crustal thickness, consideration should be given to the use of the criterion that the "best" value of the parameter is that for which the sum of the squares of the departures *from the regression line* is least, or what is equivalent, is that for which the variance of the observations *from the regression line* is least. The variances from the regression line have been calculated and are given in Table 4 and in Figure 4, for the whole country in terms of quarter-degree means and of one-degree means, and for the one-degree latitude strips using quarter-degree means. The curves in Figure 4 are similar in character to those given in the discussion of the Bowie criterion. There is again a group in which the minimum is fairly clear, another group in which the minimum is very ill-defined, and four strips, 22°, 26°, 27° and 33° S for which there is no minimum at all.

If parabolas are fitted it is found that for the quarter-degree squares and Free Air, AH20, AH30, AH40 and AH60 variances from the regression line, the minimum lies at 39·0 ± 1·6 km. For the AH20, AH30, AH40 and AH60 hypotheses

Table 4

Variances from means and variances from regression lines

Data	No. of observations	Variances (mgal)^2						
		Free Air	AH20	AH30	AH40	AH60	H114	Bouguer
Quarter-degree squares: Union of South Africa	1372	886	525	503	500	525	526	2084
		686	493	483	489	522	518	679
One-degree squares: Union of South Africa	142	501	316	303	305	342	337	2404
		412	301	298	303	342	336	429
22–23° S	38	322	184	185	183	197	202	707
		316	159	150	144	143	153	316
23–24° S	60	888	565	540	534	520	531	1019
		722	541	531	532	520	526	696
24–25° S	79	889	617	574	569	557	564	1094
		699	580	554	556	554	557	728
25–26° S	112	1393	793	773	776	797	784	1780
		1026	763	759	776	792	784	1030
26–27° S	142	727	529	510	494	484	482	1368
		566	517	507	494	477	480	479
27–28° S	160	553	434	422	424	437	430	2447
		532	426	406	397	381	389	532
28–29° S	164	705	484	470	469	489	503	2149
		678	484	469	467	478	496	675
29–30° S	127	395	247	242	257	314	283	2176
		377	240	222	220	238	232	386
30–31° S	115	530	228	218	227	284	269	2025
		507	222	217	227	272	263	509
31–32° S	112	567	217	201	198	220	223	952
		280	159	164	178	216	211	268
32–33° S	113	722	220	203	198	198	209	800
		363	160	159	166	182	176	362
33–34° S	113	748	649	640	631	623	637	1138
		700	647	636	626	614	632	679
34–35° S	37	120	186	232	270	337	302	123
		99	183	229	268	336	299	100

The upper figure in each pair is the variance from the mean; the lower, the variance from the regression line.

the minimum lies at $32{\cdot}5 \pm 2{\cdot}7$ km, and for the AH20, AH30 and AH40 hypotheses at $31{\cdot}4 \pm 2{\cdot}1$ km. For the one-degree squares the minima are at $35{\cdot}7 \pm 4{\cdot}8$ km, using Free Air to AH60, $29{\cdot}1 \pm 6{\cdot}7$ km, using AH20 to AH60, and $28{\cdot}9 \pm 7{\cdot}4$ km using AH20 to AH40 only.

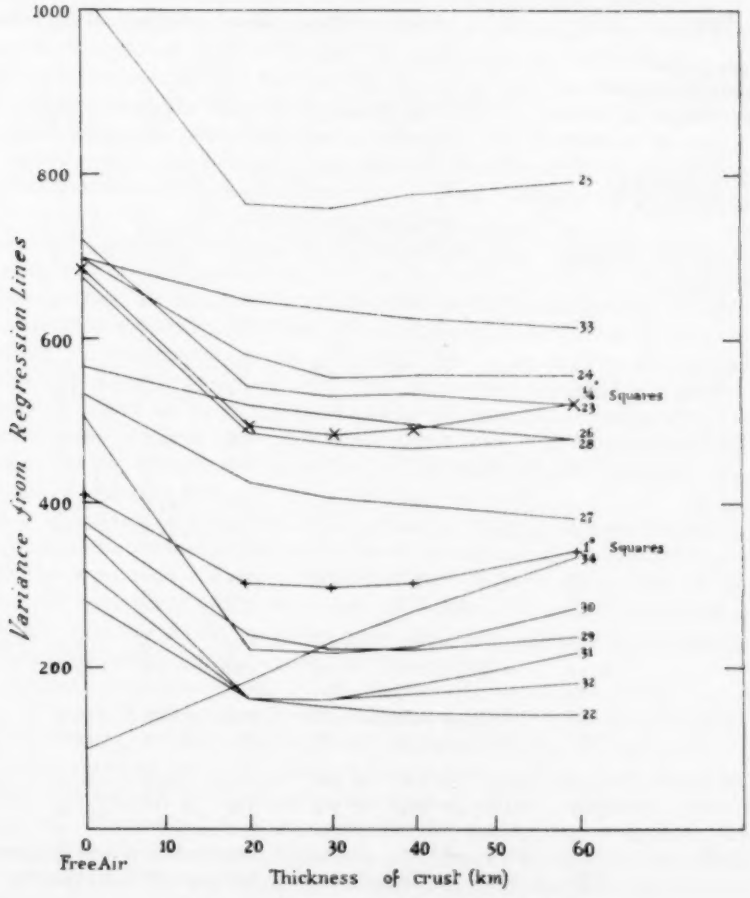

FIG. 4.—Variances from the regression lines.

5. General discussion

It is clear from the preceding sections that none of the methods used here will give estimates of crustal thickness of accuracy comparable with those which can be obtained from seismic studies of crustal structure. However, it is noteworthy that all the estimates using the Bowie criterion are between 41·8 and 33·7 km while those using the criterion of minimum variance from the regression line lie between 39·0 and 28·9 km. The higher estimates come from the parabolas fitted to the FA to AH60 data and the difference in the estimates arises in part from the fact that the curves are asymmetric. There is, moreover, no significant difference between

Table 5

Mean anomalies and standard errors of the mean

Data	Height range in feet	Mean anomalies (mgals) Free Air	AH20	AH30	AH40	AH60	H114	Bouguer
Quarter-degree squares: Union of South Africa	9	+16·0	+ 7·5	+ 1·7	− 3·7	−14·5	−13·9	− 97·6
	7057	± 0·8	± 0·6	± 0·6	± 0·6	± 0·6	± 0·6	±1·2
One-degree squares: Union of South Africa	35	+15·6	+ 7·6	+ 2·4	− 3·0	−13·4	−13·0	− 86·9
	6704	± 1·9	± 1·5	± 1·5	± 1·5	± 1·6	− 1·5	± 4·1
22–23° S	715	− 9·0	− 4·7	− 5·9	− 7·3	− 9·9	−11·2	− 77·5
	2936	± 2·9	± 2·2	± 2·2	± 2·2	± 2·3	± 2·3	± 4·3
23–24° S	806	+ 4·3	+ 1·5	− 1·5	− 4·2	− 9·5	− 9·7	− 91·0
	4542	± 3·8	± 3·1	± 3·0	± 3·0	± 2·9	± 3·0	± 4·1
24–25° S	644	−10·4	− 9·5	−13·0	−16·0	−22·7	−23·0	−113·2
	5124	± 3·4	± 2·8	± 2·7	± 2·7	± 2·7	± 2·7	± 3·7
25–26° S	83	+13·3	+ 6·5	+ 1·5	− 2·8	−11·8	−11·1	−113·0
	6636	± 3·5	± 2·7	± 2·6	± 2·6	± 2·7	± 2·6	± 4·0
26–27° S	72	+25·3	+ 9·6	+ 3·9	− 1·3	−11·5	−10·2	−121·2
	5758	± 2·3	± 1·9	± 1·9	± 1·9	± 1·8	± 1·8	± 3·1
27–28° S	112	+29·5	+18·4	+12·7	+ 7·4	− 2·9	− 2·1	−106·6
	6125	± 1·9	± 1·7	± 1·6	± 1·6	± 1·7	± 1·6	± 3·9
28–29° S	97	+16·6	+10·2	+ 4·8	− 0·3	−10·3	−10·5	−111·8
	5952	± 2·1	± 1·7	± 1·7	± 1·7	± 1·7	± 1·8	± 3·6
29–30° S	33	+21·5	+ 9·8	+ 4·0	− 1·6	−12·7	−12·2	−105·3
	7057	± 1·8	± 1·4	± 1·4	± 1·4	± 1·6	± 1·5	± 4·1
30–31° S	15	+26·4	+13·9	+ 7·1	+ 0·5	−12·4	−11·3	− 99·1
	5704	± 2·1	± 1·4	± 1·4	± 1·4	± 1·6	± 1·5	± 4·2
31–32° S	12	+36·9	+20·4	+11·8	+ 4·0	−11·3	− 8·7	− 95·8
	6344	± 2·3	± 1·4	± 1·3	± 1·3	± 1·4	± 1·4	± 2·9
32–33° S	11	+10·9	+ 3·6	− 3·3	−10·1	−23·6	−23·3	− 78·8
	5118	± 2·5	± 1·4	± 1·3	± 1·3	± 1·3	± 1·4	± 2·7
33–34° S	35	−12·1	−15·6	−22·5	−29·3	−42·6	−42·3	− 61·5
	3034	± 2·6	± 2·4	± 2·4	± 2·4	± 2·3	± 2·4	± 3·2
34–35° S	9	+14·0	+16·4	+10·8	+ 4·8	− 6·9	− 8·1	+ 0·4
	934	± 1·8	± 2·2	± 2·5	± 2·7	± 3·0	± 2·9	± 1·8

the estimates from the quarter-degree and one-degree squares. These estimates are upper limits for the crustal thickness for they are made on the basis of exact local compensation. If the compensation is regionally distributed then the crustal thickness will be less than estimated in Sections 3 and 4. It may therefore be concluded that the gravity data are consistent with a crustal thickness of 30 km and that in this region there is little evidence for regionality of compensation.

The zero regression coefficient criterion has proved to be vulnerable to systematic effects such as that found at the escarpment in South Africa.

It had been expected that the Free Air anomalies would have a greater variance from the regression line than the Bouguer anomalies on account of their correlation with small-scale uncompensated topography. It was found, however, that the variances were nearly equal. It is possible that the effect of uncompensated structure was reduced by the tendency for the observations to follow roads which tend to pass between hills rather than over them.

Another point may be noted in passing. It has been suggested by Heiskanen (1951) that in estimating the deflections of the vertical from the Stokes-Meinesz formula, it might be best to estimate mean isostatic anomalies in the sectors concerned and subsequently to reverse the mass displacements made in the isostatic reduction. The advantage of this procedure lies in the smaller variance of the isostatic anomalies as compared with Free Air anomalies. Table 4 shows that the variance about the mean is indeed less for the isostatic anomalies than for the Free Air anomalies, but only by a factor of about 0·7. It is therefore doubtful whether the use of isostatic anomalies for this purpose gives sufficient gain in precision of estimates to justify the extra labour involved.

It is of interest to note from Table 4 that the H114 hypothesis gives approximately the same variances both about the mean and about the regression line, as the AH60 hypothesis. If any considerable part of the compensation of surface topography took place within the crust, one would expect the H114 variances to be smaller than the AH60 variances, since the former hypothesis distributes the compensation between depth 114 km and the surface.

The mean anomalies for the quarter-degree and one-degree squares are given in Table 5. As has been noted earlier seismic studies of crustal structure in the Transvaal have given estimates of crustal thickness between 35 and 40 km. These estimates refer to an area with a mean height of about 1½ km. Assuming a density contrast of 0·6 g/cm^3 the thickness of the crust at sea-level would be between 27 and 32 km. The mean AH30 anomalies for the quarter-degree squares and the one-degree squares are small, indicating that gravity in South Africa does not differ greatly from that predicted on the basis of the International Gravity Formula of 1930.

6. Acknowledgments

It is with pleasure that we acknowledge our indebtedness to Dr F. Truter, Director of the Geological Survey of the Union, Dr J. F. Enslin, Dr L. E. Kent and Mr P. J. Smit for their co-operation in supplying us with the gravity observations; to Col. H. A. Baumann, former Director of the Trigonometrical Survey of the Union and Mr O. G. Reitz for co-operation in the preparation of the reduction tables. The work was begun while the senior author was on the staff of the University of Cape Town and our thanks are due to the Council of the University for a grant for clerical assistance in the reductions. Finally our thanks are due to

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Observations with the Cambridge Pendulum Apparatus in North, Central and South America in 1958

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Summary

Observations were made with the Cambridge pendulum apparatus in the period 1958 August–December at Teddington, Madison, Mexico City, Panama, Caracas, Quito, Lima, La Paz, Santiago (Chile), Buenos Aires and Rio de Janeiro. Stations were reoccupied on the return journey so that eight measures of the difference of gravity between consecutive stations were obtained. These values are reported in the paper.

1. Introduction

The idea of a series of observations with the Cambridge pendulums in South America, as a contribution to the activities of the International Geophysical Year, was discussed by Professor George P. Woollard and Mr B. C. Browne at the I.U.G.G. Conference in Toronto in 1957. Later, it was found possible to carry out this project, and the observations, which were made in the period 1958 August 13 to December 23, are reported in this paper.

The stations occupied were Teddington (National Physical Laboratory), Madison (University of Wisconsin), Mexico City (National University), Panama (Pacific Air Depot), Quito (National Observatory), Lima (Military Geographical Institute), La Paz (Military Geographical Institute), Santiago de Chile (Military Geographical Institute), Buenos Aires (Geophysics Laboratory), Rio de Janeiro (National Observatory), and Caracas (Seismological Observatory Cagigal).

Travelling was done entirely by commercial aeroplane services, a total distance of about 28000 miles, and the apparatus travelled as excess personal baggage. The two boxes of pendulums and a package containing manometers, quartz crystals and magnetometer, were carried in hand. A map of the route is seen in Figure 1.

The route was traversed in both directions, so that two visits were made to each station except Rio de Janeiro and Caracas, and three visits were made to Panama. Twenty-one sets of observations were completed.

With slight modifications, the apparatus and the observational procedure were those used by G. Jelstrup (1957) for his observations at North European stations, and by B. C. Browne in his recent work in Africa. A new Helmholtz coil, designed and constructed by L. Flavill at the Department of Geodesy and Geophysics, Cambridge, was taken to America: the frame of this coil could be separated into three 120° sections, and the coils themselves were made from very

flexible wire, so that this piece of apparatus could be packed into a box more easily portable than the one previously used. The most notable modification introduced since Jelstrup used the apparatus is that the rating of the crystal oscillator is now done by a comparison with the pulses transmitted at one-second intervals by some radio stations, such as WWV. For this reason, the crystal now used is one cut so as to oscillate at a rate very close to 100 000 c/s. The time-marking mechanism, which is operated by a synchronous motor from the 1 000 c/s output of the oscillator circuits, and which gives flashes of light at nominally half-second intervals, has been modified to include a cam disc which is rotated at 2 turns a second by the motor. On the same axis there is a free disk having a

FIG. 1.

make–break contact operated by the rotating cam, and this contact is connected in parallel with the headphones plugged to the radio receiver. The free disk can be turned by a hand-operated gear into such position that the making of the contact coincides with the receipt of each signal pulse, and so the signals are not heard. The edge of the free disc is divided by 100 markings, and it is found that the signal coincidence can be estimated to $\frac{1}{5}$ division, or 1 millisecond, with certainty. Since the timing mechanism is kept running continuously during a session of observations, for a period of some 30 to 35 hours (roughly 100 000 seconds), the accuracy of time rating is clearly more than sufficient.

The crystal used in the oscillator was "slow": its rate varied from place to place in the range 30–48 milliseconds per day, and was fairly closely correlated with the working temperature. During each session of observations, rating

checks were made several times a day and the rate was found to be practically constant at most stations: slight but measurable changes of rate at a few stations were taken into account in calculating the pendulum periods.

The transmissions from station WWV were used for time ratings, except at Teddington where a local signal was available on a wire, and at Buenos Aires where the transmission LOL was received. At the first visit to Quito, the WWV signals were very weak: otherwise the signals used were received quite easily, though at some places they could not be heard at all times of the day.

A vacuum pump operated by a 110-volt electric motor was borrowed from the University of Wisconsin: this compact lightweight instrument proved to be very useful. Several other small pieces of apparatus were borrowed from the same source.

A manometer tube broke during the journey from Madison to Mexico, and the spare manometer was used for the rest of the observations. This was tested at Buenos Aires and at Teddington, and the appropriate corrections were applied. At Mexico City, Professor Ricardo Monges Lopez very kindly arranged for two new manometers to be provided.

Local power mains, generally 110 V 60 c/s, were used to operate the crystal oscillator, radio receiver and pump at all stations except Quito, where a petrol-electric generator set was provided by the Inter-American Geodetic Survey; on this account, the observations at Quito were done in a continuous session of 24 hours, while the generator, standing in the grounds of the Observatory, was protected by a military guard.

At all stations south of Madison, the Helmholtz coils and the lamp for the recording camera were operated from batteries of three or four large dry cells: one supply of such cells, in fresh condition, just sufficed for each session of work. Accumulators were used to supply the larger currents required for the coils in the high magnetic latitudes of Madison and Teddington.

On arrival in Mexico City, I met Ingeniero Julio Monges Caldera, who accompanied me thereafter until my departure from the second visit to that city. He contributed much to the success of the project, not only through his knowledge of Spanish, but also by assisting with the observations and by attending to many details of travel and organization. I am glad to have this opportunity to acknowledge my gratitude to Julio Monges for his pleasant companionship and co-operation, without which the work would certainly have been more irksome and less expeditious. Julio Monges carried with him a Worden Gravimeter with which he made observations at the airports, the pendulum stations, and many other points in the vicinity of the places visited.

Most of the financial support for this work came from funds controlled by the Woods Hole Oceanographic Institution, and was arranged for by Professor G. P. Woollard. In close co-operation with him was Professor Ing. Ricardo Monges Lopez, Executive Vice-President of the Pan-American Institute for Geography and History, who secured the co-operation of other Latin-American countries by consultation with the Heads of the national organizations interested in Geodesy and I.G.Y. work. To these gentlemen, Col. Ferdinand J. Tate, Head of the Inter-American Geodetic Service, Ing. Col. Bolivar Zurita P., Director of the *Instituto Geografico Militar* in Quito, Col. Guillermo Barriga Meneses, Director of the *Instituto Geografico Militar* in Lima, Professor Ismael Escobar, President of the Bolivian National Committee for the I.G.Y., General Gregorio Rodriquez, President of the Executive Committee for the I.G.Y. in Santiago, General de Brigada

Victor H. J. Hoskings, Director General of the *Instituto Geografico Militar* in Buenos Aires, Professor Allyrio H. de Mattos, Director of the Division of Cartography in Rio de Janeiro, and Dr A. C. Romero, President of the National Committee for the I.G.Y. in Venezuela, we express our thanks for their support and for the personal interest they took in the work.

We are also much indebted to many other officials of the I.A.G.S. and of the various national institutions, for providing transport, and other facilities, arranging accommodation, and cheerfully spending much of their own time in helping our technical activities to proceed smoothly, in attending to various matters connected with import and export of the apparatus, and in socially entertaining us on occasions when we had free time during the visits. We would like particularly to thank Dr A. H. Cook of the National Physical Laboratory, Teddington, Dr John C. Rose of the University of Wisconsin, Professor H. Maldonado-Koerdell, Secretary of the Pan-American Committee for the I.G.Y., Dr Julian Adém of the National University of Mexico, Mr J. Koslosky of the I.A.G.S. Canal Zone, Dr Luis Eduardo Mena A., Director of the Astronomical Observatory in Quito, Dr Jorge A. Broggi, President of the Peruvian Geophysical Committee for the I.G.Y., Maj. Ing. Reynaldo Salgueiro and Capt. Salvador del Pozo of the *Instituto Geografico Militar* in La Paz, Mr Hank Hook of the I.A.G.S. in Santiago, Professor Ing. Eduardo E. Baglietto and Ing. Angel A. Cerrato of the Faculty of Engineering of the University of Buenos Aires, and Dr Lélio J. Gama, Director of the National Observatory in Rio de Janeiro.

The crystal oscillator suffered some damage on the journey from Santiago to La Paz, and Professor Escobar kindly arranged for it to be repaired by a member of his staff, to whom we are deeply indebted.

Some financial support for the project was provided by grants from the *Conselho Nacional de Geografia* and from the *Conselho Nacional de Pesquisas* in Rio de Janeiro. The visit to Caracas was not at first included in the programme but it became evident in October that time would be available for this connexion, and additional funds for the purpose were provided by the Venezuelan National Committee for the I.G.Y.

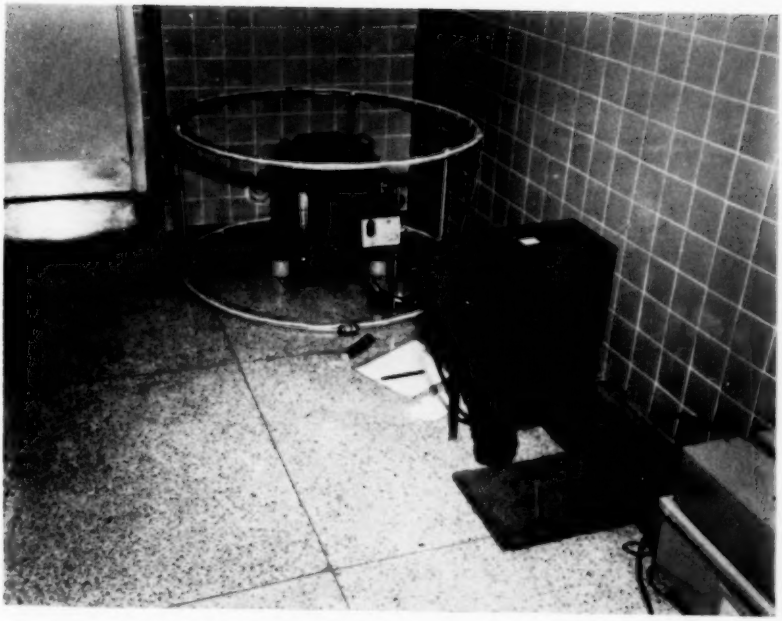
2. Stations

Full descriptions of the stations occupied are given below, and location diagrams of most of them will be seen on Figure 2. The letters RGI refer to the publication *Reseau Gravimétrique International* of the International Gravimetric Bureau.

Teddington.—Room B.17, adjoining dark room, in the basement of Bushy House on the floor about the middle of the room (The National Gravity Station in Room 11 of the Metrology Building could not be used. The value of g in Room B.17 is practically the same as at the national station.)

Madison.—Room 12A of the Geology Department adjoining lecture room at the southern end of the basement of the Science Hall, on the floor.

Mexico City.—In the new Gravitation Laboratory of the National University of Mexico, on the floor in a room adjoining the south-east corner of the large room containing concrete platforms for instruments. (The concrete platforms could not be occupied as they contain iron reinforcement and the magnetic field immediately above them is highly distorted. The value of g on the platform is



Cambridge pendulum apparatus in use.

[*Facing page* 340

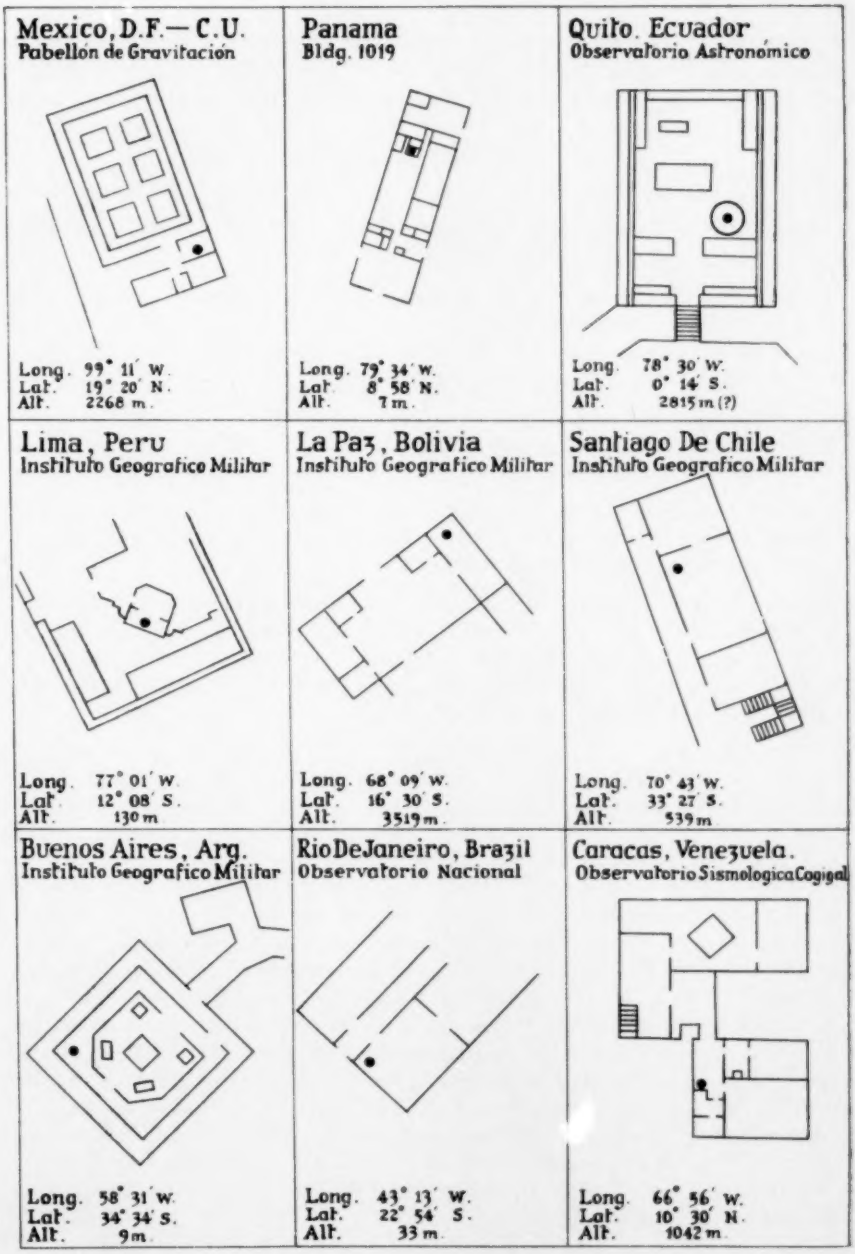


FIG. 2.—Location diagrams of South American Stations. Pendulum site: ●

about 0·1 mgal less than at the station actually occupied.) The room is shown on the diagram in RGI.

Panama.—In building 1019 on the Pacific Air Depot occupied by the I.A.G.S. Logistics Division, on the floor in a small room adjoining the radio-shop and containing a Multiplex instrument, a level calibrator and an altimeter testing table: there is a brass plug in the floor.

Caracas.—In the new Cagigal Seismological Observatory, at the end of a passage-way in the basement, on the floor outside the seismograph room.

Quito.—In the basement of the National Observatory in the city, on the floor of the seismograph room, as described in RGI.

Lima.—In the basement of the *Instituto Geografico Militar*, on the floor of a room normally used as a canteen, and close to the new photogrammetry building. (The station occupied by Iverson is in the adjoining printing shop and the difference of g between the two places is inappreciable.)

La Paz.—On the ground floor of the *Instituto Geografico Militar*, in a room used by the I.A.G.S. as a store and fitted with cupboards, on the floor about three metres from the window.

Santiago (Chile).—In the basement of a building of the *Instituto Geografico Militar*, in a room used by the I.A.G.S. as a paper store, on the floor about a metre south of the concrete wing-walls.

Buenos Aires.—In the lower basement of the building known as "*Instalaciones Sargento Major Ingeniero Jose Antonio de Condarco*", in the suburb of San Martín, on the floor in the western corner of the passage-way which surrounds the inner laboratory in which is installed the pendulum apparatus for absolute determination of gravity. Site 2a on the diagram in RGI.

Rio de Janeiro.—In a room adjoining the seismograph room on the ground floor of the National Observatory, on an isolated section of the floor specially provided to serve as a gravity reference station. Site 2 on the diagram in RGI.

3. Observations

Each set of observations consisted, as usual, of two runs lasting about seventy minutes with each of four pairs of pendulums—I.A. with I.B, I.A with I.C, VI.A with VI.B and VI.B with VI.C. The plane of swing was perpendicular to the magnetic meridian, and the vertical component of local magnetic field was neutralized by the use of the Helmholtz coils. Before each set of observations, the pendulums were tested for magnetism: in fact, they picked up very little magnetism on their travels and degaussing was only occasionally necessary. As usual, the aim was to keep to a fixed routine of observational procedure, and to have conditions, as far as possible, the same at each station. The working temperature of the pendulums varied from 14 °C at Santiago (first visit) to 26 °C at Madison (second visit). All stations were very satisfactory as regards the steadiness of the ambient temperature.

The reputed seismic activity at Santiago had obvious effects on the periods of individual pendulums there, but the means of pairs of pendulums were as consistent as at other stations.

The observed periods of the pendulums are listed in Tables 1a and 1b.

Table 1a

Pendulum periods: Set 1 (seconds)

	Outward Journey					Return journey		
	I.A	I.B	I.C			I.A	I.B	I.C
0·506	24850	25028		Teddington	0·506	24853	25046	
	24826	25060				24884	25036	
	24814		24758			24855		24755
	24804		24758			24844		24760
0·506	46214	46419		Madison	0·506	46205	46448	
	46201	46417				46201	46437	
	46192		46125			46224		46124
	46185		46114			46219		46139
0·507	09018	09257		Mexico City	0·507	09025	09251	
	09022	09240				09013	09247	
	09007		08941			09007		08956
	09009		08951			09005		08922
0·507	01246	01489		Panama	0·507	01239	01470	
	01227	01475				01220	01468	
	01236		01154			01234		01167
	01244		01147			01225		01161
				Caracas	0·507	06461	06710	
						06459	06708	
						06462		06402
						06460		06394
				Panama	0·507	01212	01485	
						01206	01454	
						01220		01163
						01221		01155
0·507	26240	26446		Quito	0·507	26221	26456	
	26239	26453				26215	26465	
	26235		26147			26230		26152
	26235		26135			26213		26150
0·507	00177	00395		Lima	0·507	00187	00403	
	00165	00388				00170	00406	
	00173		00094			00186		00105
	00162		00091			00162		00093
0·507	21338	21569		La Paz	0·507	21364	21561	
	21332	21546				21341	21562	
	21327		21248			21345		21253
	21318		21253			21337		21251
0·506	70514	70724		Santiago	0·506	70514	70730	
	70574	70664				70521	70711	
	70484		70443			70500		70425
	70522		70414			70511		70407
0·506	63349	63589		Buenos Aires	0·506	63363	63589	
	63323	63606				63363	63570	
	63326		63292			63329		63304
	63364		63272			63357		63271
0·506	86648	86891		Rio de Janeiro				
	86617	86883						
	86633		86595					
	86624		86578					

Table 1b

Pendulum periods: Set VI (seconds)

Outward journey					Return journey			
	VI.A	VI.B	VI.C			VI.A	VI.B	VI.C
0·505	46272	46356		Teddington	0·505	46321	46373	
	46274	46360				46285	46409	
		46386	46169				46413	46193
		46402	46147				46384	46218
0·505	67620	67682		Madison	0·505	67638	67729	
	67594	67673				67648	67733	
		67712	67488				67757	67548
		67707	67486				67764	67560
0·506	30351	30421		Mexico City	0·506	30339	30440	
	30344	30411				30337	30414	
		30475	30223				30483	30211
		30474	30225				40474	30211
0·506	22595	22662		Panama	0·506	22585	22644	
	22584	22660				22579	22641	
		22719	22475				22692	22474
		22711	22461				22672	22474
				Caracas	0·506	27805	27884	
						27811	27884	
							27917	27705
							27915	27701
				Panama	0·506	22584	22670	
						22581	22657	
							22707	22465
							22709	22480
0·506	47525	47607		Quito	0·506	47522	47599	
	47535	47580				47525	47613	
		47649	47402				47661	47391
		47651	47409				47652	47400
0·506	21516	21571		Lima	0·506	21494	21585	
	21514	21561				21495	21597	
		21631	21389				21640	21399
		21635	21382				21634	21385
0·506	42620	42700		La Paz	0·506	42646	42711	
	42635	42697				42627	42714	
		42748	42520				42768	42524
		42757	42518				42768	42521
0·505	91901	91941		Santiago	0·505	91905	91928	
	91901	91934				91868	91962	
		91979	91806				92016	91720
		91983	91803				92014	91735
0·505	84766	84822		Buenos Aires	0·505	84753	84783	
	84759	84837				84776	84825	
		84885	84657				84892	84626
		84868	84661				84883	84632
0·506	08034	08101		Rio de Janeiro				
	08017	08091						
		08136	07912					
		08144	07902					

Differences of acceleration of gravity were computed from changes of period between successive stations. It is not proposed to embark on an elaborate discussion of errors and accuracy because it seems more appropriate to defer any such discussion till the time comes for consideration of these results in conjunction with other contributions to the world network of gravity base stations. However, some simple indicators of internal precision may be of immediate interest. Firstly, the difference of the two determinations of mean period of a pendulum pair does not exceed 20×10^{-8}s, except in one case which is referred to below: the r.m.s. value of these differences, eighty-three of them after excluding the exceptional one, is 8×10^{-8}s (25×10^{-8}s is equivalent to a milligal). The r.m.s. of all 84 differences is $8{\cdot}7 \times 10^{-8}$s, and thus the standard deviation of the mean period from 2 pendulums is $4{\cdot}35 \times 10^{-8}$s, which is equivalent to 0·17 mgal. Hence, the s.d. of a calculated difference of gravity from one pair of pendulums is 0·24 mgal. Since the difference of gravity between each pair of consecutive stations has been measured eight times, the s.d. of the mean of the eight values is 0·09 mgal. It may be desirable, in due course, to estimate separately the s.d. of each of the ten gravity-differences measured in this series. The figures on which the foregoing discussion is based are set out in Table 2.

Table 2

Differences between the two measures of mean period of each pair of pendulums at each occupation: unit 10^{-8} *s*

	First occupation				Second occupation			
	I.AB	I.AC	VI.AB	VI.BC	I.AB	I.AC	VI.AB	VI.BC
Teddington	4	5	3	2	10	3	0	2
Madison	8	8	18	4	8	5	6	10
Mexico City	6	6	8	0	8	18	14	4
Panama (three occupations)	16	0	6	11	10	8	4	10
					18	4	8	6
Caracas (one occupation)					1	5	2	3
Quito	3	6	8	4	2	10	8	0
Lima	10	6	6	2	7	17	6	10
La Paz	14	2	6	4	10	5	8	2
Santiago	0	4	4	0	6	4	2	6
Buenos Aires	4	9	4	6	10	2	32	2
Rio de Janeiro (one occupation)	20	13	14	1				

Secondly, there were changes of mean period, from first to second occupation of the same station, as tabulated below, in units of 10^{-8} s.

Briefly, the repeat values are very satisfactory at all stations except at Madison where they are poor, and at Teddington where they are not as good as at the other nine stations. No reason for the bad repeats at Madison has come to light.

The computed differences of gravity, in milligals, are given in Table 4. Discrepancies, arising mainly from the poor repeat values at Madison, show in the results for the section Teddington–Madison–Mexico City, but mean values of outward and return results are seen to have remarkable consistency.

Table 3

Changes of mean periods of pendulum pairs, from first to second occupations at repeat stations: unit 10^{-8} *s*

	Pendulum pair			
	I.AB	I.AC	VI.AB	VI.BC
Buenos Aires	+ 4	+ 1	−12	−10
Santiago	0	− 5	− 3	−22
La Paz	+11	+10	+12	+ 9
Lima	+10	+ 6	+ 2	+ 6
Quito	− 5	− 2	+ 3	− 2
Panama	−20	− 5	− 2	− 1
Panama*	+10	+ 7	−11	−12
Mexico City	0	− 5	+ 1	− 4
Madison	+10	+22	+45	+59
Teddington	+14	+20	+32	+26

* Second to third occupation.

Table 4

Computed gravity differences (mgal)

Upper figure—outward journey. Lower figure—return journey.

	Pendulum pair			
Station	I.AB	I.AC	VI.AB	VI.BC
Teddington	− 827·9	− 827·9	− 827·5	− 827·3
	827·8	828·0	828·0	828·5
Madison	−2427·5	−2427·6	−2428·2	−2428·6
	2427·1	2426·6	2426·5	2426·1
Mexico City	+ 300·0	+ 300·2	+ 299·7	+ 299·8
	300·3	300·0	300·3	300·1
Panama	− 963·4	− 963·7	− 963·1	− 963·0
	964·0	963·8	963·2	963·0
Quito	+1005·0	+1004·8	+1004·9	+1004·9
	1004·4	1004·5	1004·9	1004·6
Lima	− 816·2	− 815·9	− 815·9	− 816·0
	816·3	816·1	816·2	816·2
La Paz	+1961·9	+1961·7	+1961·8	+1961·7
	1962·3	1962·2	1962·4	1962·9
Santiago	+ 276·5	+ 276·5	+ 275·9	+ 275·9
	276·4	276·3	276·2	275·5
Buenos Aires	− 900·2	− 900·2	− 900·5	− 900·2
	900·1	900·2	901·0	900·6
Rio de Janeiro				
Panama	− 202·4	− 202·2	− 201·8	− 201·7
	202·0	201·9	202·2	202·2
Caracas				

The other results are much more satisfactory. Repeat values are good, particularly at Mexico City, and there is little sign of any systematic change in the pendulums. Mention must be made, however, of the observations with pendulum pair VI.AB at Buenos Aires on the second visit: in this case the difference of the two measures of mean period had the exceptionally large value of 38×10^{-8}s, referred to above. It may be advisable to ignore this observation and regard the sum of the affected gravity differences, −624·8mgal, as a measure of the difference of gravity between Rio de Janeiro and Santiago.

In conclusion, I would like to express my indebtedness to Mr B. C. Browne for much help and encouragement in carrying out this work.

Department of Geodesy and Geophysics,
Cambridge.
1959 *August* 7.

Reference

G. Jelstrup, 1957. *Observations on the gravimetric calibration base.* Geografiske oppmåling, Oslo.

Wave Velocities below the Mohorovičić Discontinuity*

B. Gutenberg

(Received 1959 August

Summary

The velocities of longitudinal waves immediately below the Mohorovičić discontinuity which have been found by various authors for continental regions decrease on the average from about 8·2 km/s, if the Mohorovičić discontinuity is at a depth of 30 km, to about 8·0 km/s, if it is at a depth of 50 km. The rate of this decrease in velocity of 0·011 ± 0·0016 km/s per km is statistically significant and exceeds appreciably the critical rate of about 0·0013 km/s per km, required for a low-velocity channel. The corresponding calculated average decrease in the velocity of transverse waves from about 4·64 km/s, if the Mohorovičić discontinuity is at a depth of 30 km, to about 4·56 km/s, if it is at 50 km (at a rate of 0·004 ± 0·0056 km/s per km) can be expected to represent average conditions under continents only if this decrease in velocity is considered in connection with that for longitudinal waves. For a depth of about 60 km, both calculated velocity-depth curves connect smoothly with those found by the author from the apparent velocities at the points of inflection of travel-time curves for earthquakes originating at various depths. The new results confirm the hypothesis that under continents the asthenosphere low-velocity channel starts at the Mohorovičić discontinuity.

Various seismologists have calculated the depth of the Mohorovičić discontinuity and the wave velocities immediately below it on the basis of travel-time curves found from supposedly refracted waves recorded after nearby earthquakes or artificial explosions. A number of such data reported during the past ten years for continental areas are given in Table 1. A plot of these velocities in Figure 1 shows that the resulting velocities V for longitudinal waves decrease definitely with increasing depth h of the Mohorovičić discontinuity, while the corresponding velocities v of the transverse waves show a less definite decrease with increasing h.

If we assume that the velocities for both wave types change linearly with the depth h in the range between $h = 26$ km and $h = 50$ km, for which data are available, the method of least squares gives the following relationships:

$$V = (8{\cdot}08 \pm 0{\cdot}013) - (0{\cdot}011 \pm 0{\cdot}0016)(h - 40)\ \text{km/s}, \quad (1)$$

$$v = (4{\cdot}60 \pm 0{\cdot}044) - (0{\cdot}004 \pm 0{\cdot}0056)(h - 40)\ \text{km/s}. \quad (2)$$

* Contribution No. 939 from the Division of the Geological Sciences, California Institute of Technology.

Table 1

Velocities V of longitudinal and v of transverse waves and depth h of the Mohorovičić discontinuity reported for various continental regions. A and E in the last column indicate whether the results are based on artificial explosions or on earthquakes respectively.

Region	Author	h km	V km/s	v km/s	A,E
NW Germany	Reich, Förtsch, Schulze (1951)	26	8·32	4·58	A
NW Germany	Reinhardt (1954)	27	8·19	—	A
NW Germany	Willmore (1949)	28	8·18	—	A
Bavarian Alps	Reich (1958)	28	8·2	—	A
Black Forest	Rothé (1958)	30	8·2	4·4	A
SW Germany	Hiller (1953)	30	8·1	4·7	E
Bohemia	Kárník (1956)	31	8·15	—	A
Alaska	Aldrich & others (1957)	31	8·1	—	A
Po Valley	Caloi (1958)	32	8·16	4·52	E
Kyoto	Kishimoto (1955)	32	8·1	4·7	E
U.S.A., Atlantic coast	Adams & others (1952)	33	8·1	—	A
S. Africa	Gane & others (1956)	34	8·22	4·70	A
Pennsylvania, New York	Katz (1953)	34	8·14	4·69	A
Hungary	Vanek (1955)	35	8·14	4·53	E
SE of Lake Balkash	Gamburtsev, Veytsman (1956)	35	8·1	—	A
S. California	Gutenberg (1952)	35	8·15	4·5	A
West Australia	Bolt & others (1958)	35	8·21	4·75	A
Canadian Shield	Hodgson (1953)	36	8·18	4·85	A
East Transvaal	Hales, Sacks (1959)	37	7·96	—	E
West Australia	Doyle (1957)	37	8·18	4·72	A
NW of Lake Issik Kul	Gamburtsev, Veytsman (1956)	38	8·1	—	A
NE of Lake Balkash	Gamburtsev, Veytsman (1956)	40	8·1	—	A
Wisconsin	Slichter (1951)	42	8·17	—	A
Central Appalachians	Adams & others (1952)	45	8·06	—	A
N. of Lake Issik Kul	Gamburtsev, Veytsman (1956)	45	8·1	—	A
S. Alps	Caloi (1958)	45	8·0	4·41	E
NE India	Anonymous (1955)	46	7·91	4·46	E
Caucasus depression	Balavadze, Tvaldvadze (1957)	49	8·0	4·65	A
Central Asia	Bune, Butovskaya (1955)	50	8·0	—	A
Chilean Andes	Aldrich & others (1958)	50±	8·0±	—	A

Both lines are shown in the respective portions of Figure 1. While the rate of velocity decrease with h for longitudinal waves is statistically significant, the numerical result for the corresponding rate for transverse waves is not. However, the fact that its trend is similar to that for longitudinal waves makes it very probable that the rate of decrease in the transverse velocity given by equation (2) is not far from correct.

The deviations of the observed values from the calculated straight lines have various causes. First, the basic data of Table 1 are affected by errors. The waves on which the calculated velocities are based are rather small; as a consequence of the low-velocity channel below the discontinuity, they are diffracted. In several instances, different authors have arrived at somewhat different velocities on the basis of the same records (for example, for the Heligoland explosion, first three lines of Table 1). Frequently, this is a consequence of different interpretation of the recorded phases. Tilt of the Mohorovičić discontinuity, effects of variable structure of the Earth's crust, and differences in the material below the

discontinuity may produce additional errors. Moreover, the assumption of linear equations connecting the velocities and the depth h is only an approximation.

The greater scattering for v than for V results partly from the fact that longitudinal waves refracted below the Mohorovičić discontinuity are the first waves in seismograms recorded beyond a certain distance from the source, while transverse waves always arrive at times when the motion from earlier waves has not completely subsided. Moreover, frequently no clear transverse waves are recorded from artificial explosions.

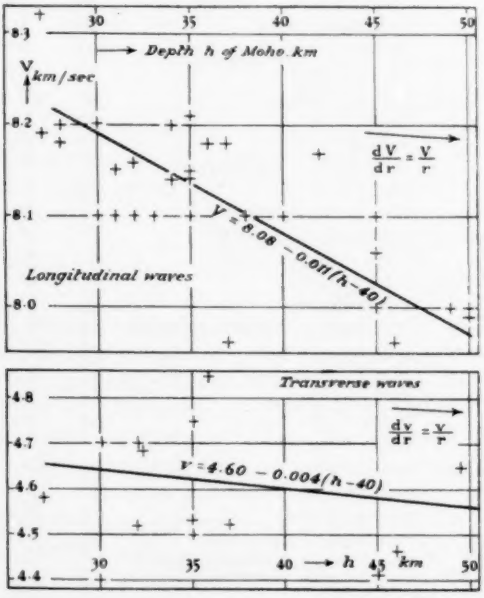


FIG. 1.—Velocities V of longitudinal waves and v of transverse waves observed in continental areas below the Mohorovičić discontinuity as function of its depth h in km.

Results from explosions in oceanic regions cannot be combined with results obtained in continental areas. First, there is a greater probability of differences in material below the Mohorovičić discontinuity. However, even if the material is the same, the differences in temperatures at a given depth under oceans and under continents may produce noticeable differences in wave velocities; the effect of differences in pressure is probably smaller. Data from ten different localities in the Pacific Ocean with depths h of the discontinuity between 9 and 20 km (average 12·7) indicate velocities V of between 7·9 and 8·4 km/s (average 8·16) while similar data from the Atlantic Ocean with h between 10 and 14 km (average 12 km) give values for V between 7·9 and 8·1 km/s (average 7·98 km/s). Similarly, in New Zealand values of 7·9 to 8·1 km/s have been found for V and about 4·5 km/s for v corresponding to $h = 20$ km (Eiby 1958).

Extrapolations of the equations (1) and (2) indicate that V and v decrease to about 7·75 and 4·48 km/s respectively at a depth of 70 km. Unpublished values found by the author by use of the apparent velocity at the point of inflection of travel-time curves for earthquakes at intermediate depth (compare Gutenberg

1953) give about 7·8 and 4·44 km/s respectively for depths of about 80 km. These new velocity–depth curves for depths greater than 60 km merge into the lines given by Equations (1) and (2) without marked discontinuity. Representative values for V and v as a function of depth down to 400 km are listed in Table 2. It does not seem likely that the final values resulting from a revision for depths greater than 60 km, now under way, will result at any depth in changes in excess of 0·1 km/s for longitudinal waves and of 0·05 km/s for S.

Table 2

Representative values of the velocity V of longitudinal waves and v of transverse waves in km/s and of Poisson's constant σ in the upper portion of the Earth's mantle under continents as a function of the depth d in km
(*July* 1959)

d	40	50	60	70	80	100	150	200	250	300	350	400
V	8·08	7·97	7·87	7·82	7·80	7·82	7·95	8·1	8·3	8·55	8·75	9·0
v	4·60	4·55	4·51	4·47	4·44	4·42	4·40	4·43	4·53	4·65	4·78	4·93
σ	0·26	0·26	0·26	0·26	0·26	0·265	0·28	0·29	0·29	0·29	0·29	0·29

On the basis of the new results there can be little doubt that the asthenosphere low-velocity channel starts for both wave types immediately at the Mohorovičić discontinuity. It also appears that the differences in velocities found in various continental regions immediately below the Mohorovičić discontinuity depend mainly on the depth of the discontinuity and, except perhaps for areas near the oceans, not so much on local conditions below the crust. A detailed description of the increasing number of other methods and phenomena which support the hypothesis of the asthenosphere low-velocity channel has been given by Gutenberg (1959, pp. 75–89).

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California Institute of Technology,
Pasadena, California.
1959 *July*.

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A Palaeomagnetic Study of some Lower Jurassic Rocks of N.W. Europe

R. W. Girdler

(Received 1959 January 8)

Summary

A study has been made of the directions of natural remanent magnetization of the Upper Lias Sands of England (Section 1) and reasons are given for considering the magnetization of some of these to be stable. The results show that the direction of the Earth's magnetic field in Lower Jurassic times was significantly different from that of the present Earth's field, having an easterly declination and inclination 64° to 70°.

The work was extended (Section 2) to include some Lower Jurassic volcanic rocks from Southern France and these gave results in general agreement with those for England.

It is concluded (Section 3) that the ancient geomagnetic pole position for the Lower Jurassic period lies in the region of the Caspian Basin.

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So far, few palaeomagnetic results have been obtained for rocks of Jurassic age from Great Britain. Nairn (1956, 1957) has given results for Jurassic material collected from North Yorkshire and Scotland. The estuarine sediments of Yorkshire have directions of magnetization close to the present Earth's field and Nairn (1957) gives reasons for considering these to be unstable. From Scotland, three limestone samples from the base of the Estuarine Beds of Skye and one sandstone sample from the Brora Arenaceous Series were found to be reversely magnetized.

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The age relationships were determined by S. S. Buckman (1889, 1910) and are summarized in Figure 1. The mineralogy of the sands has been described by P. G. H. Boswell (1924).

It was concluded that the sands are suitable for palaeomagnetic work for the following reasons:

(a) They occur over a wide area and are well dated.

(b) The uniformity and fineness of grade suggest that they were deposited in quiet water, free from violent current and wave action.

(c) The sands have remained fresh and unaffected by subsequent geological events and it is possible that they have an undisturbed "depositional" remanent magnetization.

The stratigraphical horizons sampled are shown approximately in Figure 1 by the letters *N* or *R*, the letters referring to horizons found to have normal or reverse directions of magnetization respectively.

Two or three specimens were cut from each sample. The specimens were in the form of cylinders with height equal to diameter (equal to 3·5 cm) and a special perspex holder was constructed so that a cylinder could be measured in six positions on-centre beneath the astatic magnetometer. The six positions correspond to placing the six faces of a cube in turn beneath the magnet system. The direction

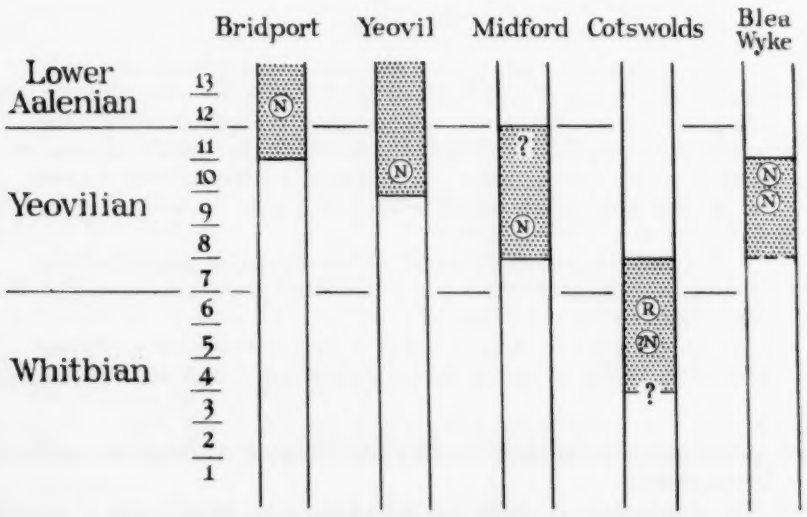


FIG. 1.—Age relationships of the Upper Lias diachronous sands (adapted from Arkell 1933). The numbers refer to the ammonite subzones given by Dr W. J. Arkell.

of magnetization was computed from the sums ΣX, ΣY, ΣZ of the eight deflexions for the x, y, z directions of the orthogonal co-ordinate reference system chosen for the cylinder.

The directions of natural remanent magnetization for all the specimens measured are plotted on Schmidt equal area projections. Figures 2 and 3 show the directions of magnetization of the Bridport and Yeovil Sands respectively. The directions for the Midford Sands are shown in Figures 4a and 4b; for the Midford Sands it was necessary to correct for the tilt of the strata and Figures 4a and 4b show the results before and after applying the geological corrections. Figures 5a and 5b show the directions for the Cotswold Sands; some of these are reversely magnetized (Figure 5a) and some are normally magnetized (Figure 5b). The Blea Wyke Beds of Yorkshire consist of grey, micaceous sandstone (38 ft) overlain by soft, yellow sandstone (30 ft); the directions of magnetization of these are shown separately in Figures 6a and 6b respectively. The mean directions for all the results are illustrated in Figure 7.

It is evident from Figure 7 that the Upper Lias Sands show varying degrees of stability of magnetization. The Yeovil Sands and the Blea Wyke grey sands are magnetized in directions close to the Earth's present field and are presumed to be unstable; the normally magnetized Cotswold Sands, the Blea Wyke yellow sands and the Bridport Sands have directions of magnetization ranging from 6° to 23° to the east of the Earth's present field and it is suggested that these are partially stable; the reversely magnetized Cotswold Sands and the normally magnetized Midford Sands have a natural remanent magnetization in an east–westerly direction and it is likely that these are stable.

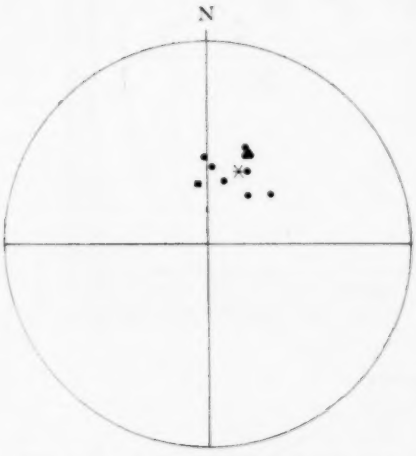


FIG. 2.—Directions of natural remanent magnetization of the Bridport Sands.

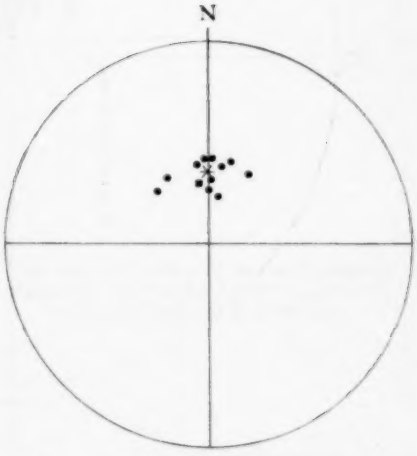


FIG. 3.—Directions of natural remanent magnetization for the Yeovil Sands.

In this and succeeding diagrams the plane of the projection (Schmidt equal area) is the equatorial plane; normal directions of magnetization (downwards) are shown by dots and reverse directions of magnetization (upwards) are shown by circles; the mean direction is shown by an asterisk and the direction of the Earth's present magnetic field by a square.

The following reasons are given for considering the reversely magnetized Cotswold Sands and the normally magnetized Midford Sands to be stable:

(a) The occurrence of a reverse direction of magnetization is in itself evidence of stability.

(b) Figures 4a and 4b show that the directions of magnetization for the Midford Sands become less scattered after geological corrections have been applied (Graham's stability test, 1949). This suggests that the magnetization was acquired before tilting and that the direction of polarization has since been preserved.

(c) The mean direction of magnetization for the normally magnetized Midford Sands is slightly further from the Earth's present field than the axis of magnetization of the reversely magnetized Cotswold Sands suggesting that the Midford Sands are at least as stable as the reversely magnetized Cotswold Sands.

(d) Several specimens were re-measured after a period of six months after having been stored in random directions in the Earth's field No significant changes in the directions of natural remanent magnetization were observed.

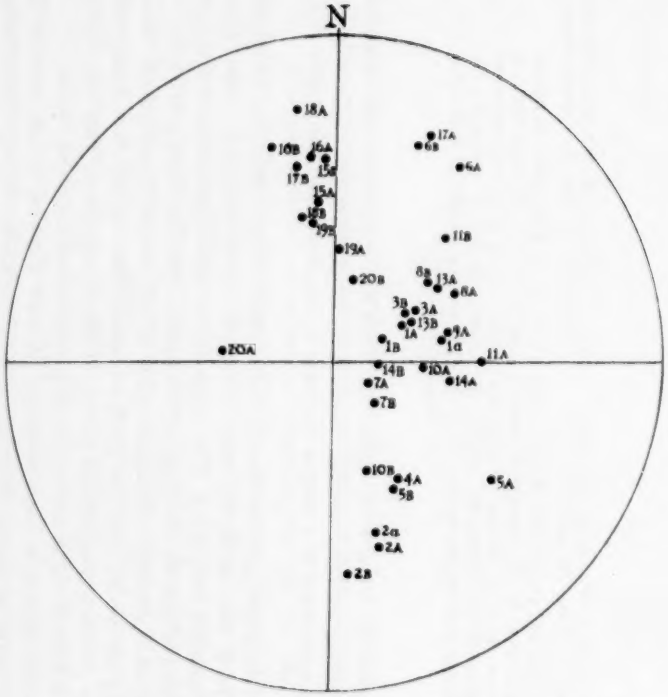


FIG. 4a.—Directions of natural remanent magnetization for the Midford Sands.

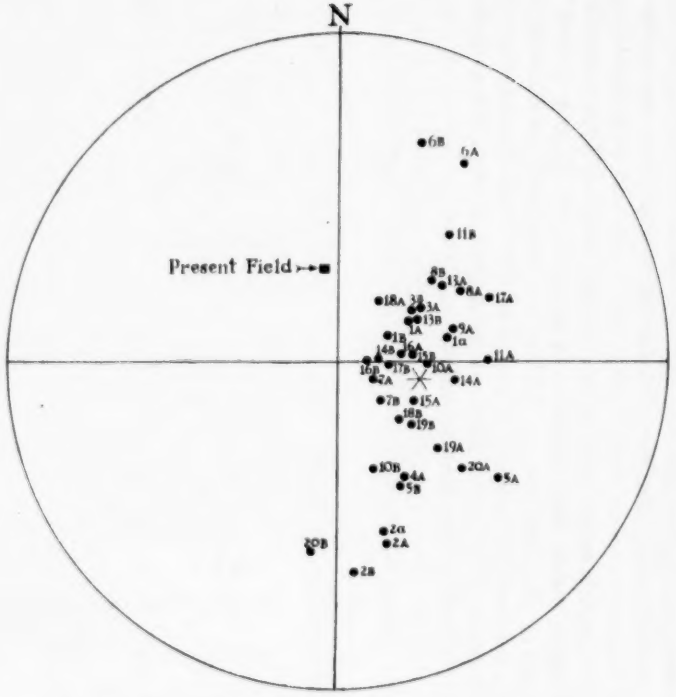


FIG. 4b.—Directions of natural remanent magnetization for the Midford Sands after correction for geological dip.

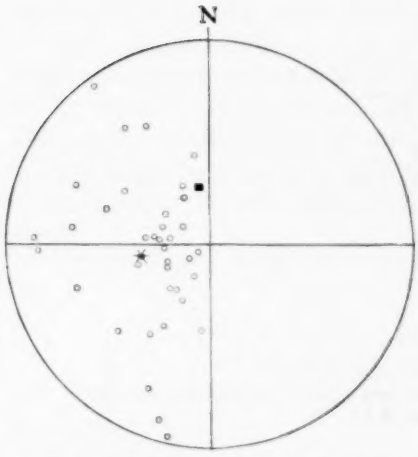


FIG. 5a.—Directions of natural remanent magnetization for the Cotswold Sands—reversely magnetized specimens.

FIG. 5b.—Directions of natural remanent magnetization for the Cotswold Sands—normally magnetized specimens.

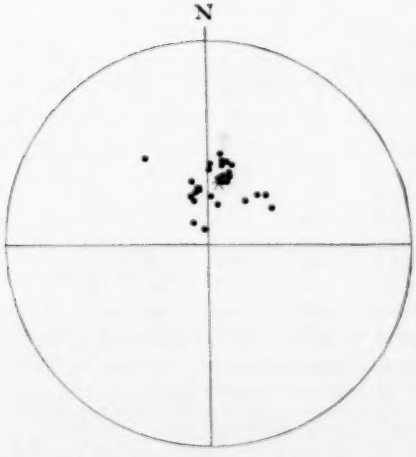
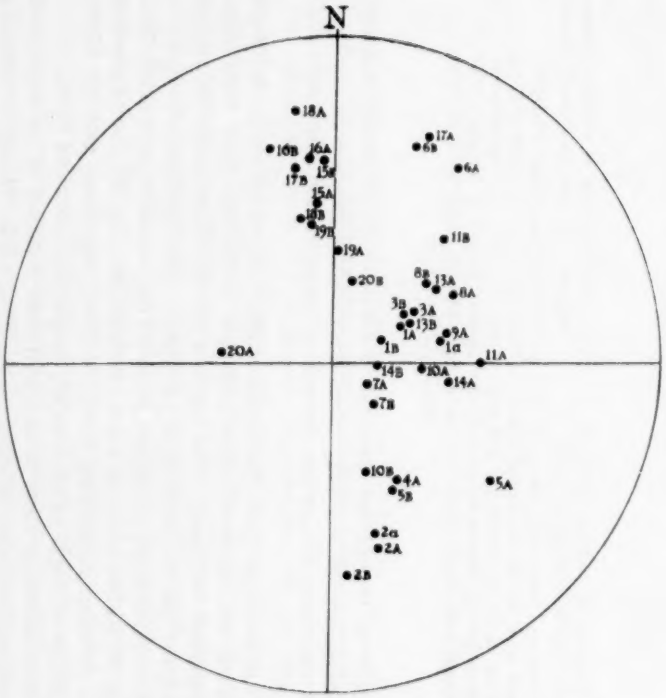


FIG. 6a.—Directions of natural remanent magnetization for the Blea Wyke Beds—yellow sands.

FIG. 6b.—Directions of natural remanent magnetization for the Blea Wyke Beds—grey sands.



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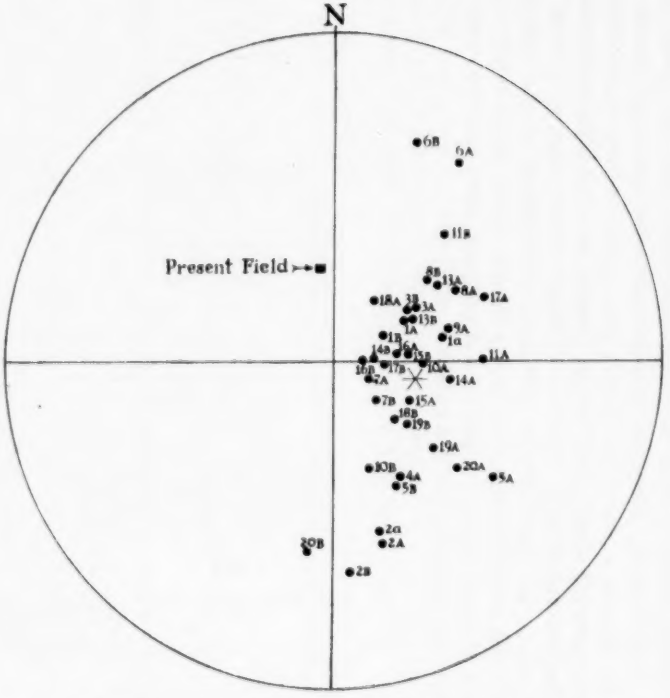

FIG. 4b.—Directions of natural remanent magnetization for the Midford Sands after correction for geological dip.

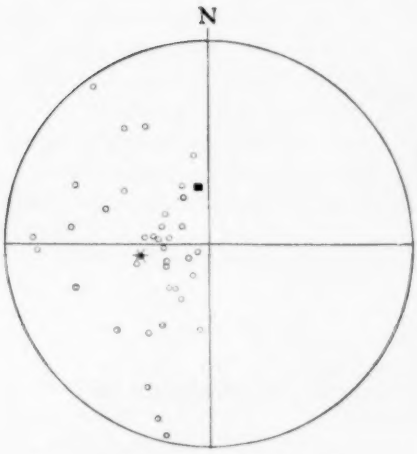

FIG. 5a.—Directions of natural remanent magnetization for the Cotswold Sands—reversely magnetized specimens.

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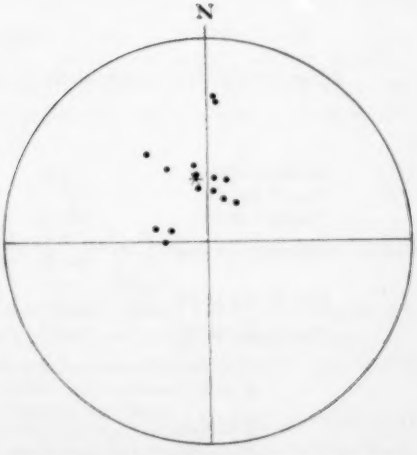


FIG. 6b.—Directions of natural remanent magnetization for the Blea Wyke Beds—grey sands.

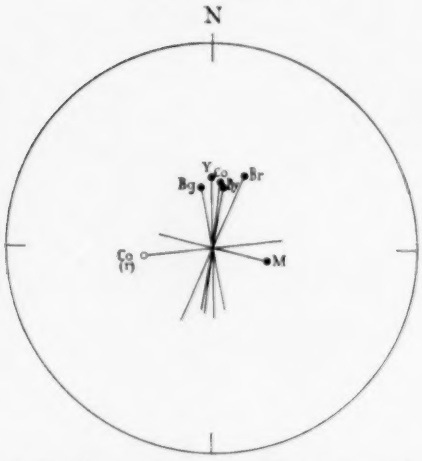


FIG. 7.—Mean directions of natural remanent magnetization for the Upper Lias Sands.

Br = Bridport Sands.
Y = Yeovil Sands.
M = Midford Sands.
Co = Cotswold Sands—normally magnetized samples.
Co(r) = Cotswold Sands—reversely magnetized samples.
By = Blea Wyke Beds—yellow sands.
Bg = Blea Wyke Beds—grey sands.

The values for the mean directions of natural remanent magnetization and the Fisher statistics (Fisher 1953) are given in Table 1.

Table 1

Mean directions of magnetization for the Upper Lias sands of England

		θ °	ϕ °	α °	K	N	S
Bridport Sands		23·4	60·4	5·1	73·1	12	7
Yeovil Sands		358·9	63·9	6·1	57·3	11	10
Midford Sands		103·5	69·8	7·4	10·2	42	20
Cotswold Sands	N	6·5	65·5	5·5	22·1	32	17
	R	262·6	−64·1	9·9	6·5	38	17
Blea Wyke Beds	Yellow	9·7	67·0	4·2	46·0	26	9
	Grey	349·2	66·3	9·1	18·4	15	5
Hettangian tuffs		54·9	58·9	5·8	17·0	38	18

θ = Declination, measured clockwise from north.
ϕ = Inclination, with reference to the horizontal plane.
(positive inclination—downwards negative inclination—upwards.)
α = Radius of 95 per cent circle of confidence (Fisher 1953).
K = Fisher's precision factor.
N = Number of cylinders measured.
S = Number of rock samples.

Figures 4b and 5a show that the normally magnetized Midford Sands and the reversely magnetized Cotswold Sands have a larger scatter in the directions of magnetization For this reason, the sampling was extended and the mean directions of magnetization have been calculated from 42 specimens (20 samples) for

the Midford Sands and 38 specimens (17 samples) for the reversely magnetized Cotswold Sands. The following are possible reasons for the large scatter:

(a) It is possible that some of the sands have been disturbed due to movement of the underlying Upper Lias Clays.

(b) In some cases, there is a possibility that the sands have been disturbed by vegetation, especially where collection was made from sunken lanes.

(c) The fine grade of the sands and the absence of any obvious bedding planes makes it difficult to ascertain accurate geological corrections.

The results show that some Lower Jurassic sediments of Great Britain have a mean direction of polarization with declination N. 83–103° and inclination 64–70°.

Table 2 shows a comparison with Nairn's (1957) results.

Table 2

Comparison with Nairn's results

	θ °	ϕ °	α °	K	N	S
Nairn (1957)	234·4	−66	6·6	33·5	14	4
Midford Sands	283·5	−69·8	7·4	10·2	42	20
Cotswold Sands	262·6	−64·1	9·9	6·5	38	17

(The symbols have the same meaning as in Table 1)

The main differences are in the values for the declination which are 28° to 49° larger for the Cotswold and Midford Sands than for Nairn's rocks.

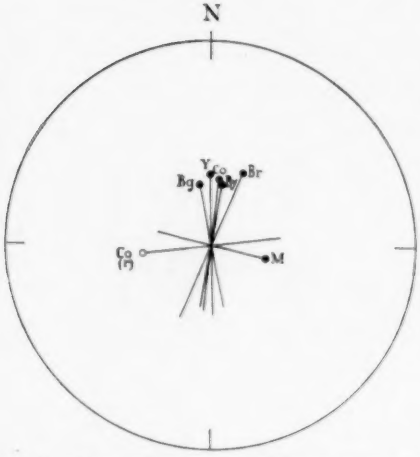
Nairn's specimens are mainly from the Middle Jurassic and are therefore slightly younger in age than the Upper Lias Sands. However, the comparison may not be justified as Nairn's results depend on measurements from only four samples and no stability tests were applied.

2. Directions of magnetization of some Lower Jurassic Volcanic rocks from France

A collection has been made of some basalts and tuffs of Lower Hettangian age from the northern foothills of the Pyrenees, the object being to compare the directions of natural remanent magnetization with those obtained for the Lower Jurassic sedimentary rocks of England.

Results are given for volcanic rocks collected from an area west of Foix (Latitude 42°57′ N, Longitude 1°35′ E). In this area volcanic tuffs occur in the middle of the Hettangian Limestone and are therefore well dated. The tuffs contain fragments of basalts, limestones and marls and are of variable texture. The thickness varies from 1·5 to 10 m. Stratigraphical details have been given by Dubar (1925).

The volcanic rocks of the northern Pyrenees are equivalent in age to the Lower Lias of Great Britain. The stratigraphical relations with respect to the Upper Lias Sands discussed in Section 1 are shown in Table 3.



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Co = Cotswold Sands—normally magnetized samples.
Co(r) = Cotswold Sands—reversely magnetized samples.
By = Blea Wyke Beds—yellow sands.
Bg = Blea Wyke Beds—grey sands.

The values for the mean directions of natural remanent magnetization and the Fisher statistics (Fisher 1953) are given in Table 1.

Table 1

Mean directions of magnetization for the Upper Lias sands of England

		θ	ϕ	α	K	N	S
		°	°	°			
Bridport Sands		23·4	60·4	5·1	73·1	12	7
Yeovil Sands		358·9	63·9	6·1	57·3	11	10
Midford Sands		103·5	69·8	7·4	10·2	42	20
Cotswold Sands	N	6·5	65·5	5·5	22·1	32	17
	R	262·6	−64·1	9·9	6·5	38	17
Blea Wyke Beds	Yellow	9·7	67·0	4·2	46·0	26	9
	Grey	349·2	66·3	9·1	18·4	15	5
Hettangian tuffs		54·9	58·9	5·8	17·0	38	18

θ = Declination, measured clockwise from north.
ϕ = Inclination, with reference to the horizontal plane.
(positive inclination—downwards negative inclination—upwards.)
α = Radius of 95 per cent circle of confidence (Fisher 1953).
K = Fisher's precision factor.
N = Number of cylinders measured.
S = Number of rock samples.

Figures 4b and 5a show that the normally magnetized Midford Sands and the reversely magnetized Cotswold Sands have a larger scatter in the directions of magnetization For this reason, the sampling was extended and the mean directions of magnetization have been calculated from 42 specimens (20 samples) for

the Midford Sands and 38 specimens (17 samples) for the reversely magnetized Cotswold Sands. The following are possible reasons for the large scatter:

(a) It is possible that some of the sands have been disturbed due to movement of the underlying Upper Lias Clays.

(b) In some cases, there is a possibility that the sands have been disturbed by vegetation, especially where collection was made from sunken lanes.

(c) The fine grade of the sands and the absence of any obvious bedding planes makes it difficult to ascertain accurate geological corrections.

The results show that some Lower Jurassic sediments of Great Britain have a mean direction of polarization with declination N. 83–103° and inclination 64–70°.

Table 2 shows a comparison with Nairn's (1957) results.

Table 2

Comparison with Nairn's results

	θ °	ϕ °	α °	K	N	S
Nairn (1957)	234·4	−66	6·6	33·5	14	4
Midford Sands	283·5	−69·8	7·4	10·2	42	20
Cotswold Sands	262·6	−64·1	9·9	6·5	38	17

(The symbols have the same meaning as in Table 1)

The main differences are in the values for the declination which are 28° to 49° larger for the Cotswold and Midford Sands than for Nairn's rocks.

Nairn's specimens are mainly from the Middle Jurassic and are therefore slightly younger in age than the Upper Lias Sands. However, the comparison may not be justified as Nairn's results depend on measurements from only four samples and no stability tests were applied.

2. Directions of magnetization of some Lower Jurassic Volcanic rocks from France

A collection has been made of some basalts and tuffs of Lower Hettangian age from the northern foothills of the Pyrenees, the object being to compare the directions of natural remanent magnetization with those obtained for the Lower Jurassic sedimentary rocks of England.

Results are given for volcanic rocks collected from an area west of Foix (Latitude 42°57′ N, Longitude 1°35′ E). In this area volcanic tuffs occur in the middle of the Hettangian Limestone and are therefore well dated. The tuffs contain fragments of basalts, limestones and marls and are of variable texture. The thickness varies from 1·5 to 10 m. Stratigraphical details have been given by Dubar (1925).

The volcanic rocks of the northern Pyrenees are equivalent in age to the Lower Lias of Great Britain. The stratigraphical relations with respect to the Upper Lias Sands discussed in Section 1 are shown in Table 3.

Table 3

Age relations of Upper Lias sands and Lower Lias volcanic rocks of Pyrenees

Stage	Formation	Facies
Toarcian	Upper Lias	Bridport Sands Yeovil Sands Midford Sands Cotswold Sands
Upper Pliensbachian	Middle Lias	
Lower Pliensbachian Sinemurian Hettangian	Lower Lias	Volcanic rocks of north Pyrenees

As the whole of the Jurassic material was laid down in approximately 40 million years, from 140 to 180 million years ago (Mayne & others 1959) the maximum age difference between the Hettangian tuffs and the Upper Lias Sands is likely to be less than 12 million years.

Eighteen orientated samples were collected from five sites west of Foix. Unfortunately, no time was available for geological mapping which was necessary to ascertain accurate corrections for the dip of the strata. The geological corrections have been estimated with the aid of field notes and survey maps and these have been applied to the directions of natural remanent magnetization before computation of the mean directions of magnetization given in Table 1. There is likely to be little error in the values used for geological strike but it is possible that some of the corrections for geological dip may be in error by $\pm 15°$.

Specimens were cut in the form of cylinders and measurements made as described in Section 1.

The mean directions of magnetization together with the Fisher statistics are given in Table 4. The directions for all the samples are illustrated in Figures 8a and 8b: Figure 8a shows the computed directions and Figure 8b the directions after the geological corrections have been applied.

Table 4

Mean directions of magnetization for the Lower Lias volcanic rocks of France

	θ	ϕ	α	K	N	S
	°	°	°			
Hettangian tuffs	54·9	58·9	5·8	17·0	38	18

The rocks are considered to have a stable magnetization for the following reasons:

(a) Nearly all the specimens are magnetized in a south-easterly direction (Figure 8a), far from the direction of the Earth's present field.

(b) When geological corrections are applied the directions of magnetization become less scattered (Figure 8b) and this constitutes an application of the stability test of J. W. Graham (1949).

FIG. 8a.—Directions of natural remanent magnetization for Hettangian tuffs from the northern Pyrenees Central area.

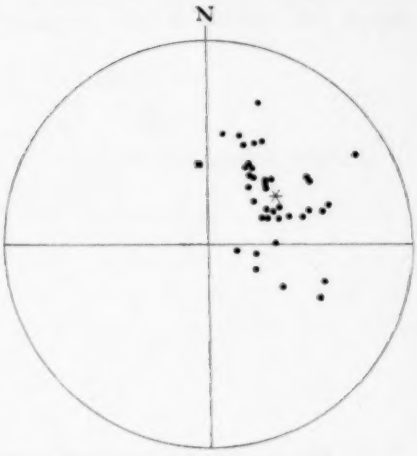

FIG. 8b.—Directions of natural remanent magnetization for Hettangian tuffs from the northern Pyrenees Central area after geological corrections have been applied.

3. Geomagnetic pole positions for the Lower Jurassic period

Pole positions have been computed from the directions of natural remanent magnetization of the Midford Sands, the reversely magnetized Cotswold Sands and the Hettangian volcanic rocks. The lengths of the semi-major and semi-minor axes of the confidence ovals have been calculated from the Fisher radius of confidence at the 5 per cent level by the method suggested by Creer & others (1957). These are given in Table 5 and illustrated on a stereographic projection for the northern hemisphere in Figure 9.

Table 5

Pole positions for the Lower Jurassic

	Pole position		Confidence Oval	
	Lat.	Long.		
Midford Sands	32°·8N	41°·0E	12°·8	11°·0
Cotswold Sands	38·2	59·3	15·7	12·5
Pyrenees Central area	49·0	76·6	8·6	6·4

It is seen from Figure 9 that the pole positions computed from two completely different rock types from different areas show reasonably good agreement and it is concluded that the geomagnetic pole for the Lower Jurassic period as calculated from rocks from Great Britain and France lies in the general region of the Caspian Basin.

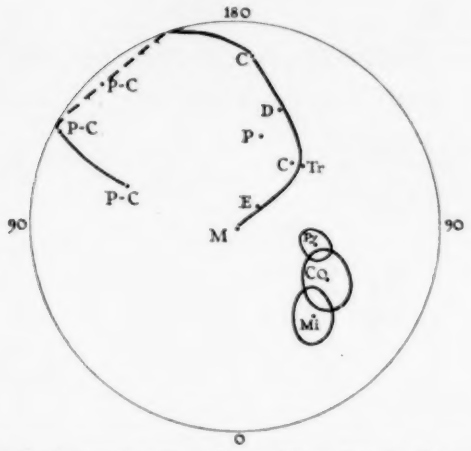

FIG. 9.—Ancient geomagnetic pole positions for the Lower Jurassic period.

Mi = Pole position computed from mean direction of magnetization of the Midford Sands.
Co = Pole position computed from mean direction of magnetization of reversely magnetized Cotswold Sands.
Py = Pole position computed from mean direction of magnetization of Hettangian tuffs from the Pyrenees Central area.

The polar wander curve given by Creer & others (1957) is also shown. The symbols in stratigraphical order are: M = Mioccene, E = Eocene, Tr = Trias, P = Permian, C = Carboniferous, D = Devonian, C = Cambrian, P-C = Pre-Cambrian.

4. Acknowledgments

The work reported on above was carried out at the Department of Geodesy and Geophysics of the University of Cambridge.

My sincere thanks are due to:

Mr B. C. Browne, Mr J. C. Belshé, Dr A. E. M. Nairn (King's College, University of Durham) and Mr R. Wilson (Imperial College, London) for helpful discussions.

Dr D. V. Agar (Royal School of Mines, London), Mr E. Cameron (University of Manchester), Dr D. T. Donovan (University of Bristol) and Dr G. A. Kellaway (H.M. Geological Survey) for map references where suitable exposures of the Upper Lias Sands were found.

Mr J. C. Belshé, Mr D. R. Reeves, Mr G. A. Day and Mr M. J. Keen for their help on rock collecting expeditions.

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Research Note

A Note on the Relation between Wind Velocity and the Amplitude of Microseisms on the coast of Antarctica

J. MacDowall

(Received 1959 June 29)

Summary

Considerable microseismic activity was observed at the Royal Society Base, Halley Bay during the summer season from December to February, particularly during on-shore winds. Examples of microseismic and wind observations from 1957 December 30 to 1958 March 6 are presented, to illustrate the nature of the relation between the two observations, and to show the rapid reduction in microseismic activity near the end of February.

1. Introduction

The amplitude of microseisms with a one second period recorded on the vertical component of a Willmore seismograph was measured at the Royal Society Base, Halley Bay (75°31′ S 26°37′ W) from 1957 December 25 to 1958 May 10. During a brief summer season from December to February considerable microseismic activity was observed which interfered with the records in wind speeds greater than about ten knots.

2. The site

The position of the Royal Society Base with respect to the ice-front is illustrated by Figure 1. It was on an ice-shelf which fringed the east coast of the Weddell Sea. The greater part of the ice-shelf was floating but local grounding occurred in places near the ice-front.

In a 30 ft pit the seismometers were installed about 75 ft above sea-level. The low level of microseismic disturbance in winter was thought to confirm that this site was free from mechanical influences due to the base, but the records were occasionally subject to electrical interference.

3. The seismograph and methods of measurement

The seismograph had maximum sensitivity near one second and was equipped with quarter-second galvanometers. When used on full sensitivity, the amplitude sensitivity was about 0·07 microns per mm.

The amplitude and period of the microseisms were measured as recommended in "IGY Instructions for Seismological Stations" but a unit of 10^{-2} microns was used.

Microseisms were predominantly of 1·0 s period but at times of intense activity a component with 0·5 s period was seen. On several occasions waves with a period of from 15 s to 30 s were observed which lagged the onset of shorter period waves by 15 hours or more. The magnification for 20 s period waves was about one-tenth of that for 1 s waves. In this note only the common 1 s microseisms were considered.

The vertical component only was measured. Horizontal movement was of the same magnitude and approximately proportional to the vertical movement.

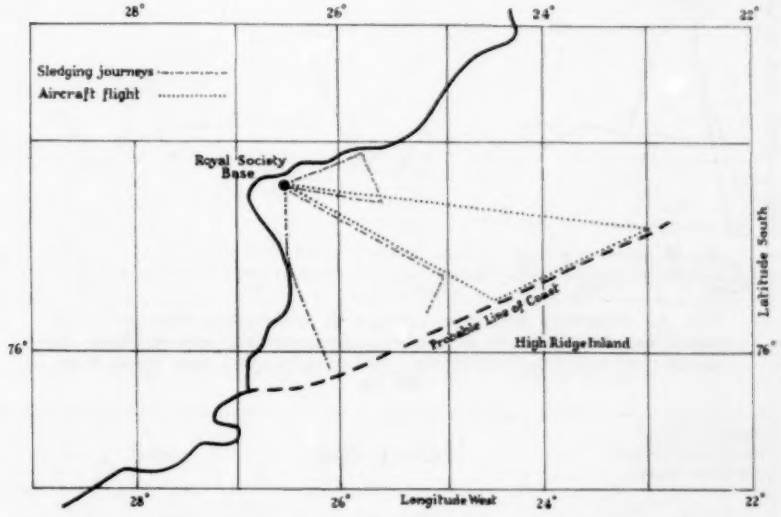


FIG. 1.—Ice front and position of Royal Society Base taken from Admiralty Chart 3176.

(Mercator projection 1:2 750 000 at Lat. 68 °.)

4. Discussion of the observations

In Figures 2 to 4 the amplitude of microseisms was plotted at three-hourly intervals together with the wind observed at six-hourly intervals. The close association between wind speed and the amplitude of microseisms was seen from December to February. Wind direction, too, affected the microseisms, notably on January 22 when a strong east wind had less effect on the amplitude than a moderate or fresh north-west wind on January 21 or the fresh south-west wind on January 18. The lesser effect of wind from the land on February 7 to February 11 is also noted in Figure 3.

Near the end of February, however, a marked change occurred in the relationship between wind and microseisms (Figure 3), so by the first week of March a strong south-west wind had only small effect on microseismic activity (Figure 4).

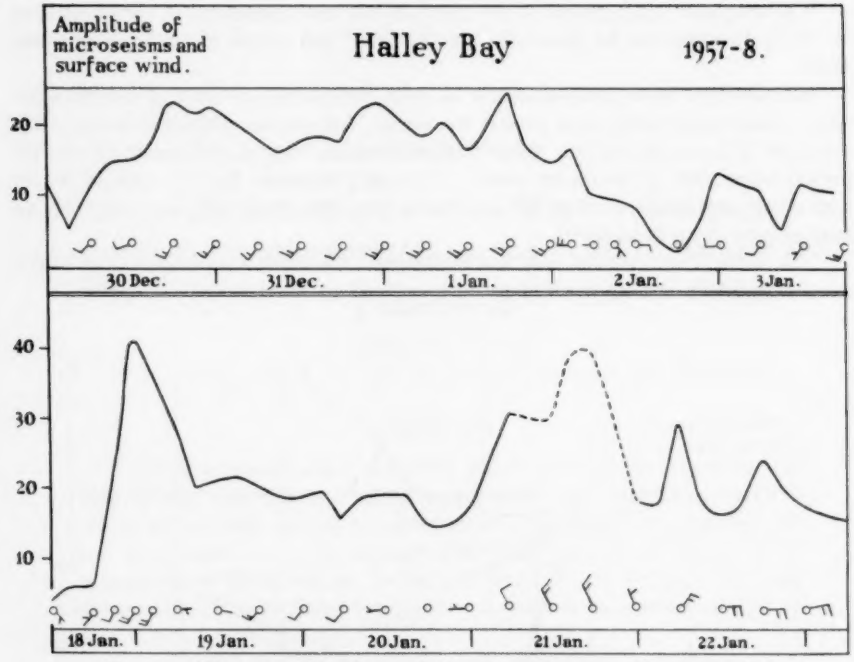


FIG. 2.—Illustrates the close relation in midsummer between wind speed and the amplitude of one second period microseisms. Note on January 22 how an off-shore wind had a lesser effect than winds from the sea.

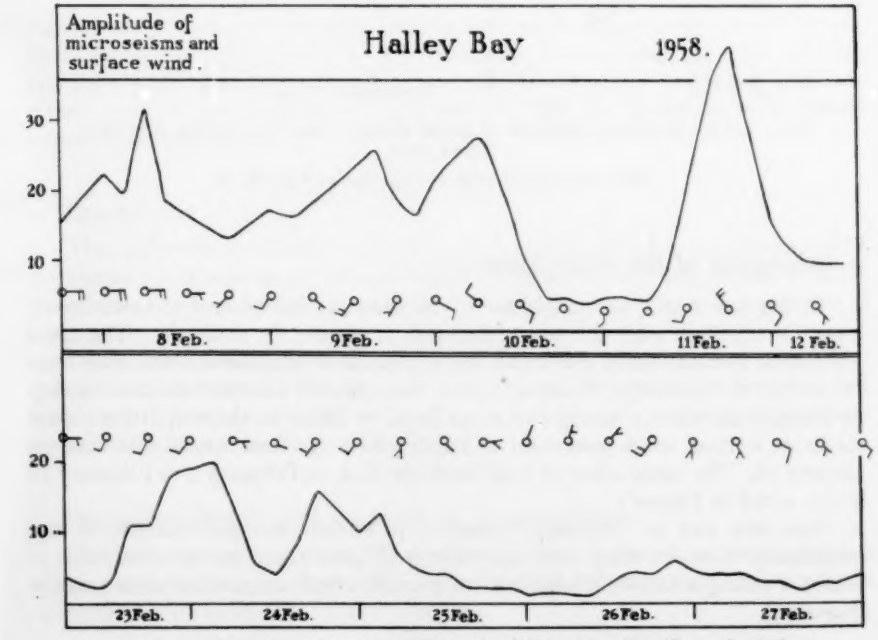


FIG. 3.—This shows how near the end of February the effect of wind on microseisms became less.

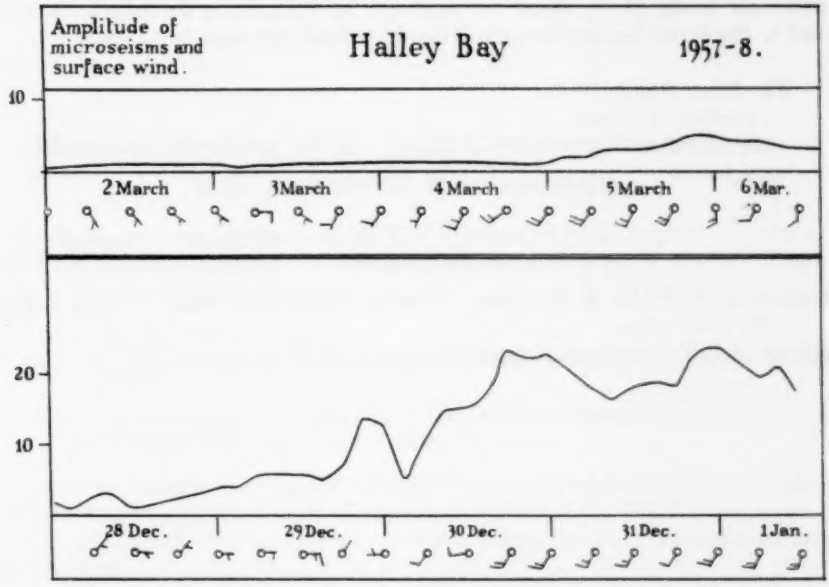
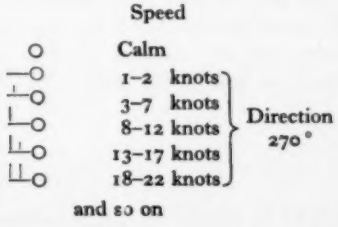


FIG. 4.—Contrasts the greatly reduced effect of wind in early March with the effect of wind on microseismic activity at the end of December in similar winds.

Note.—In Figures 2–4 above, the arrows point along the direction of the wind. In general a whole feather indicates a speed of 10 knots as follows:



5. Conclusion

At Halley Bay microseisms were only active during three summer months, particularly during on-shore winds. For the remaining eight or nine months microseisms became active only during prolonged spells of the strongest winds. It was therefore concluded that the microseisms originated at the ice-front and that the cover of sea-ice damped out this movement for three-quarters of the year and stifled microseismic activity.

6. Acknowledgments

The work was performed at the Royal Society base while the author was seconded from the Meteorological Office. He is indebted to his colleagues, in

particular to Mr D. G. Ward, for assistance in maintaining the seismic records, and to the Royal Society for permission to publish this note.

The Royal Society,
Burlington House,
*London, W.*1:
1959 *June* 26.

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Moscow Meeting of the Joint Commission on Solar and Terrestrial Relationships

The Commission met at the University of Moscow on 1958 August 12, prior to the tenth General Assembly of the International Astronomical Union. There were three parts to the meeting:

(i) A short business meeting.
(ii) A colloquium comprising individual communications on solar–terrestrial relations.
(iii) A joint discussion (that is, jointly with Solar Commissions of the I.A.U.) on "Flares and Corpuscular Streams".

Of these the business meeting and colloquium are reported below, while the joint discussion will be reported in the *Transactions of the I.A.U.*, Vol. X.

In 1958 March, five months before the meeting, the Joint Commission had circulated its

Neuvième Rapport de la Commission pour l'étude des Relations Entre les Phénomènes Solaires et Terrestres.

This is a comprehensive report setting out developments in solar–terrestrial relations during the years 1954–57. It contains review articles by W. O. Roberts, M. A. Ellison, J. Rösch, R. Tousey, H. Friedman, F. S. Johnson, R. Michard, P. Simon, W. R. Piggott, C. W. Allen, V. C. A. Ferraro, M. Koomen, J. Heidmann, and P. Bernard. 572 copies have been distributed to persons and institutions known to be interested (285 to I.U.G.G. members, 168 to I.A.U. members, 103 to U.R.S.I. members and 16 others). Some copies are still available by application to Professor C. W. Allen, University of London Observatory, Mill Hill Park, London, N.W.7.

Business meeting

The Commission assembled at the University of Moscow on 1958 August 12, and was attended by:

L. d'Azambuja (President)	delegates of the I.A.U.
M. A. Ellison	
G. Righini	
J. Rösch	
E. Vassy	delegates of the I.U.G.G.
M. Nicolet	
C. W. Allen (Secretary)	
J. F. Denisse	delegates of U.R.S.I.
D. H. Menzel	

The President began by welcoming the new members of the Commission who were present, Professor D. H. Menzel who had been a Councillor and Dr J. F. Denisse. An apology for the absence of Professor J. Bartels was presented by Professor Vassy.

The President explained to the meeting that the future constitution of the Mixed Commission was hanging in the balance. Various proposals had been put forward for the reorganization of the Commission. There appeared to be three possibilities:

(1) That the Mixed Commission should be suppressed.
(2) That it should continue its work, possibly with a modification of structure.
(3) That it should be absorbed into the I.A.U.

He explained that the triennial Reports of the Commission were much appreciated and that these should be retained if possible. Distribution amounted to over 500 copies. If the Commission were absorbed into the I.A.U. then these reports would in future be reduced to no more than I.A.U. Commission reports and would thus lose the comprehensiveness and individuality that has been their principal appeal.

The views of the meeting were then asked for.

Professor M. Nicolet considered that if the Mixed Commission were absorbed into the I.A.U. both the value and the spirit of the Commission would be lost. He suggested that the Commission, with its present independent status, should be modified and improved. Possibly there should be a Directing Committee of about seven members. Members present strongly supported the view that this Commission should be associated *jointly* between the I.A.U. and I.U.G.G. and U.R.S.I.

Dr d'Azambuja communicated with regret his resignation as President, owing to his impending retirement from official participation in solar work.

He proposed the name of Professor C. W. Allen as President and Dr M. A. Ellison as Secretary. This proposal was seconded by Professor Menzel and carried.

On the proposal of Professor Allen a vote of thanks was passed to Dr d'Azambuja for his long and distinguished services to the Commission and for his work in connection with the publication of its Reports.

The Commission unanimously proposed that Professor S. Chapman, the foundation member who had been first President (1924–1934) should be named Honorary President of the Commission, and that Dr G. Abetti, President 1934–1946, and Dr L. d'Azambuja, President 1946–1958, should continue their part in the Commission under the title Honorary Members.

Membership at the close of the Meeting, August 1958

Members from I.A.U.: M. A. Ellison (Secretary), G. Righini, M. Waldmeier.
Councillors from I.A.U.: S. B. Nicholson, J. Rösch.
Members from I.U.G.G.: H. Alfvén, J. Bartels, E. Vassy.
Councillors from I.U.G.G.: M. Hasegawa, M. Nicolet, O. R. Wulf.
Members from U.R.S.I.: C. W. Allen (President), A. E. Covington, J. F. Denisse, D. H. Menzel.
Honorary President: S. Chapman.
Honorary Members: G. Abetti, L. d'Azambuja.

Colloquium on Solar–Terrestrial Relations

The colloquium was open for contributions on any matters concerned with solar–terrestrial relations. Those contributions that have been or will be adequately published elsewhere are here reported in title only.

Sur Quelques Relations Entres les Phénomènes Solaires et Ceux de la Basse Atmosphère

E. Vassy

On considère quelques phénomènes de la basse atmosphère qui paraissent liés à l'activité solaire:

Polarisation de la lumière solaire diffusée: corrélation entre la position des points neutres et l'activité solaire (*Busch*).

Scintillation et ondulation des étoiles variant avec les aurores et les perturbations magnétiques (*Bigourdan*).

Indice de réfraction de l'air lié à l'activité solaire (*Tilton*).

Formation de cirrus à la suite des aurores polaires (*Tromholt, Dauvillier*).

Observations pyranométriques au voisinage du Soleil (*Abbot*).

Luminance, polarisation et couleur de la lumière solaire diffusée au zénith dépendant de l'activité magnétique (*Barber*).

Phénomène de *G. Piccardi.*

Couleur de la Lune éclipsée (*Danjon*).

Lumière cendrée de la Lune (*Dubois*).

On envisage ensuite le mécanisme possible de l'action solaire. Pour cela on s'attache d'abord à expliquer la discontinuité de la luminance de la Lune éclipsée au voisinage du minimum d'activité solaire. En effet pour être efficaces les taches ou jets coronaux doivent se trouver le plus près possible de l'équateur solaire. Or la latitude moyenne des taches part de 30° au début du cycle pour arriver vers 8° au minimum suivant. On examine la nature des rayonnements responsables et la voie dans laquelle doivent s'engager les recherches.

Discussion

J. Houtgast: Down to what level in the Earth's atmosphere can we expect the solar particles to penetrate—10 km or below?

F. Link: A la question de M. Houtgast relative au siège des variations dans l'atmosphère terrestre qui influence la luminosité des eclipses de Lune on peut dire a la base de la théorie des éclipses que ce sont les regions *basses* (troposphère) qui entre dans la loi de Danjon, car cette loi donne des variations vers le centre de l'ombre terrestre.

On the Scale of importance of Solar Flares

M. C. Ballario

[Described by G. Righini]

This communication deals with the general results deduced from photometric measurements of Hα solar flares obtained at the Arcetri Observatory, and with a comparison between the evaluation of importance of flares, when the adopted criterium is:

(a) the intensity of the perturbed region;
(b) the area of the same region.

The material which has been used are the spectroheliograms obtained daily at the Arcetri solar tower in Hα radiation. The solar image has a diameter of 64 mm. The spectroheliograms are calibrated with a step-filter, using the continuum spectrum in the neighbourhood of Hα.

(a) Photometric measurements have been made with a Hartmann microdensitometer in which the region explored in the plate has been reduced to a disk of about 0·3 mm.

From these measurements the intensity of the most brilliant points of the perturbed regions have been obtained. This intensity is compared with that of the non-perturbed region nearby. The ratio of the two intensities is denoted by r.

The importance of the perturbed regions has been established with visual estimation, controlled, when possible, with the data of the Quarterly Bulletin and other similar publications.

The comparison between photometric data and visual importance is given in the histograms of Figure 1. The statistic refers to 350 phenomena: chromospheric faculae and flares.

As the agreement between visual estimation and photometric data is rather good it seems convenient to establish photometric limits for the various classes of importance. In this way the estimations of importance have a more objective base. These limits are deduced from the histograms and the adopted scale is:

$\text{Log } r < 0{\cdot}35$	sub-flare
$0{\cdot}35 \leqslant \text{Log } r \leqslant 0{\cdot}48$	flare of importance 1
$0{\cdot}49 \leqslant \text{Log } r \leqslant 0{\cdot}60$	,, ,, ,, 2
$0{\cdot}61 \leqslant \text{Log } r$	,, ,, ,, 3

If the intensity of a flare is referred to the continuum spectrum at the centre of the disk, a central facula is a flare when its intensity rises to $\geqslant 0{\cdot}43$ of the continuum.

(b) The area of about 200 flares, of which photometric importance was previously established, has been measured. The adopted criteria are the following.

(i) Only the most brilliant regions with well-defined contours have been measured.

(ii) If the perturbed region is constituted of two or more brilliant points, the area is the sum of the areas of the single points.

(iii) The measurements have been made directly on the spectroheliograms with a graticule and with the aid of an eye-piece.

(iv) When the light-curve of a flare was known, the area has been measured at the maximum of intensity.

The observed area has been multiplied by sec h (h = heliocentric angle of the flare) and the corrected area has been deduced. To avoid the error introduced by this factor, phenomena not too near the limb have been generally selected.

The measurements have been performed by two independent observers: the average deviation from the mean, for the projected area, is $\pm 0{\cdot}25$ square degrees.

The 210 flares have been divided in five groups according to their area. The limits, for sub-flares and flares of various importance, are those adopted for the A.G.I.:

Importance	1 −	1	2	3	3 +
Corrected area in square degrees	2·06	2·06–5·15	5·15–12·4	12·4–24·7	24·7

In addition, each flare has been classified according to the photometric scale given in Section (a) above. The histograms of Figure 2 give, for each group of

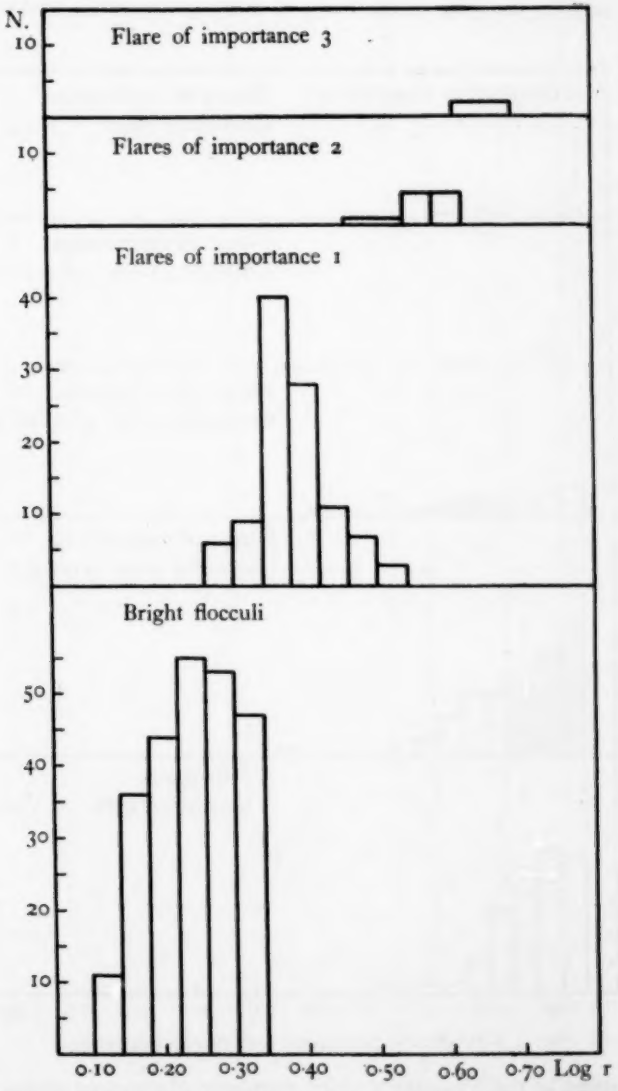

FIG. 1.—Abscissae: the logarithm of the measured intensity. Ordinates: the number of phenomena, respectively of visual importance 0, 1, 2, 3, which have the intensity included in constant intervals.

areal importance, the number of phenomena whose intensities are contained in an interval $\Delta \log r = 0{\cdot}05$. The dotted lines show the limits of the photometric importance.

It appears quite clear that the correlation between the areal importance and the photometric importance is very poor.

Furthermore one sees that:

(a) the relative intensity of the flares of every group varies within very large limits.

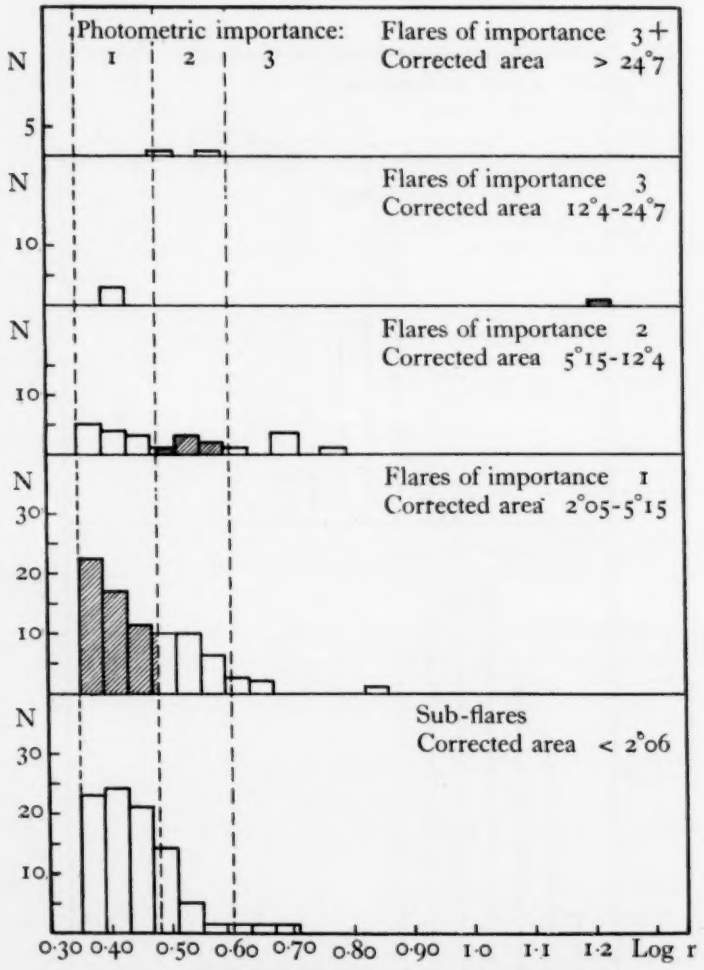

FIG. 2.—Brightness histograms for varying area ranges.

(b) the zones of the histograms which represent phenomena whose areal importance is in agreement with the photometric one (shaded in the diagram) contain only 27 per cent of all phenomena.

(c) the disagreement is more remarkable for the sub-flares which often have a large intensity. This last result supports the decision taken by the CSAGI to classify the sub-flares of great intensity as flares of importance 1, and to introduce the intermediate classes 1+ and 2+.

Figure 3 gives the relation between log r and the observed area in square

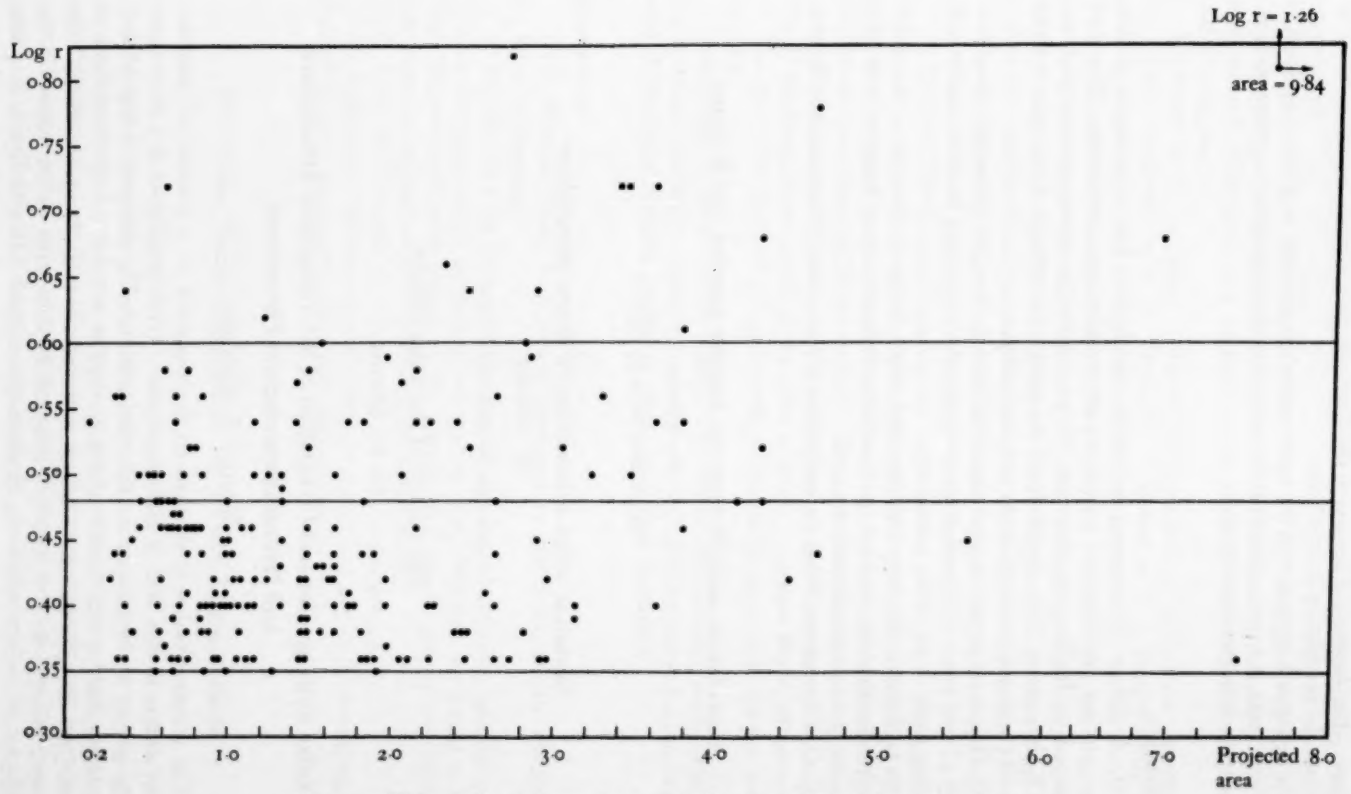

FIG. 3.—The relation between log r and the observed area in square degrees.

degrees. The horizontal lines give the limits of photometric importance; also in this case the correlation is very poor.

As perturbed regions with the same extension can have very different intensity and vice versa, it appears clear that the most consistent classification must contemplate both extension and intensity.

Discussion

M. A. Ellison: Our present system of classification had to be based on flare areas at a time when we had inadequate information about intensities. The work described by Righini, together with the photometric measures coming from the I.G.Y. flare patrol films, may indicate the need for a revision of the system which will take into account both areas and intensities.

Y. Öhman: I would suggest that Commission II might investigate the possibility of two types of classification—a scale of 1 to 5 based on areas and a scale running from − to + for intensities.

W. O. Roberts: We have experimented with a rather crude index = Area × intensity × duration, but we have not found that it correlates any better with S.I.D.s than the present classification by area.

K. O. Kiepenheuer: What we really need is a measure of the integrated brightness over the whole region.

Correlation analysis between sunspot number and K-index

K. Toman

(see *Nature*, **161**, 642, 1958)

Certaines effets solaires sur la haute atmosphère

M. Nicolet

(see *Ann. Géophys.*, **15**, 1, 1959)

The nature of the solar corona

D. H. Menzel

Solar Activity, General Circulation of the Terrestrial Atmosphere and Hydrometeorological Phenomena

Maurice S. Eigenson

The communication is devoted to the summary of a number of post-war Soviet studies in this field of Heliogeophysics. (Heliogeophysics is a short name of the science of the solar–terrestrial relationships.) The summary was published in a much more comprehensive form in 1957 in the Lvov University Press as "Essays on the physical-geographical manifestations of the solar activity" (in Russian) and will be also issued in enlarged form in 1959 in Leningrad (as the "Problems of Heliogeophysics", Hydrometeorological Editorial Board, in press, in Russian).

In the present abstract only the principal post-war Soviet results can be enumerated.

The main result is the comprehensive proof of the very close relation of the intensity and type of the general circulation of the terrestrial atmosphere with the solar activity. This proof is a collective result of many Soviet scientific institutes and investigators.

Especially useful for solar-circulation problems are the new indices introduced for (a) air circulation, by L. A. Vitels, N. A. Belinsky, G. I. Vanghengeim and B. L. Dzerdseijevsky and for (b) solar activity, by the writer.

For example as was shown quite recently by N. I. Tiabin the coefficients of correlations of some especially important hydrometeorological phenomena with our new solar index S_m (for the northern solar hemisphere) is equal to 0·99. [S_m is the mean maximum area for one sunspot group in the given apparition.]

The usefulness of the new solar index is apparently based on its peculiarity as an empirical index of geoactive *corpuscular* solar radiation.

In the many Soviet investigations the various solar cycles (with the durations of 5–6 years, 11 years, 20–50 years, 80–90 years and some centuries) and the 27 days solar rotation period in their relations to different hydrometeorological phenomena were thoroughly investigated. The most definite results were obtained in the establishment of many hydrometeorological manifestations of the *secular* (i.e. of the 80–90 years) cycle of the solar activity.

The new solar cycle of 5–6 years was preliminarily established by the writer in 1953.

New proofs in favour of the reality, importance and probable solar origin of the so-called Bruckner's cycle of hydrometeorological phenomena (i.e. the cycle with the duration between 20 and 50 years) have been produced.

Discussion

W. O. Roberts: I have been greatly interested by this review of solar effects on the lower atmosphere.

At Boulder we have been investigating the relationship between specific magnetic storms and weather changes on succeeding days. We find, for example, that when there is a pressure trough at the 300 mb level this is greatly amplified in cases where a magnetic storm occurred 2 days previously.

F. Link: What length have you found for the secular period?

M. S. Eigenson: Maximov's work gives 90 years, that is 7 or 8 11-year cycles, depending upon the lengths of the individual cycles.

The notion "Radio Importance" as a Flare Characteristic

A. D. Fokker

It is well known that a certain number of flares, especially the greater ones, are associated with outbursts at radio frequencies. Flares may be associated with bursts which are restricted to the metre- or to the decimetre-band, or with "universal" bursts, which cover the whole frequency range. The relation between flares and outbursts is of a very complicated nature. Three types of flare-associated radio events by now have been recognized.

The radio behaviour of a solar flare represents an aspect which to a large extent is independent of the optical properties. Studying the flare-associated radio phenomena one feels the need of a simple procedure, by which an importance rating, which summarizes the strength of the overall radio response to a flare, can be assigned. This importance figure should be the radio analogue to the optical flare importance.

As a basis for the definition of radio importance we have adopted the radio responses at 200, 550 and 3000 Mc/s. At these frequencies 24-hour information is available since the beginning of the I.G.Y. Each of these frequencies can be considered as representative for part of the radio spectrum of outbursts. The presence and strength of a radio response at each of these frequencies has been taken into account by giving it a certain weight. The radio importance of a particular radio event is determined by the sum of the weighting figures which are contributed by the responses at individual frequencies. An outburst which is present at one frequency only with moderate strength receives importance 1. An outburst which occurs at all three frequencies with great strength receives importance 3+. The intermediate importances are defined according to a simple rule. We are fully aware of the over-simplification which is brought about by the adoption of a straightforward rule. However, when treating a great number of cases and when trying to find some general trends, one cannot but take recourse to some simplifying procedure. Maybe the result will prove to be of a very limited value, but there is no *a priori* reason why the notion of radio importance would be less fruitful than the notion of optical importance.

The merit of radio importance can be investigated with special regard to whether or not the flare gives rise to a sudden ionospheric disturbance. It turns out that the SID efficiency of flares increases steadily with radio importance. The following figures were found:

	radio imp. 0	r. imp. 1	r. imp. 2	r. imp. 3
Proportion SID–ass. flares/total	0·06	0·29	0·48	0·90

Considering only flares which produced a radio response, we found the following mean values of importance:

	mean radio imp.	mean optical imp.
Flares without SID	1·20	1·03
SID-associated flares	1·62	1·51

Of all SID's 57 per cent were found to be accompanied by a radio event. For individual flares, though, all combinations of optical and radio importance may occur. The two characteristics, optical imp. (O) and radio imp. (R) can be considered as components of what could be named the "strength" of a flare. The strength, S, could for example be defined as: $S^2 = R^2 + O^2$.

Whether a particular flare produces a radio response or not seems to depend on certain circumstances, which as yet are unknown. It has appeared that for certain sequences of flares, which originate in closely the same region, these circumstances evidently are remarkably constant. We discovered two series of flares, which during one or two days distinguished themselves by either an abnormally small or by a very great production of radio responses. For this reason one might consider these flares as to belong to a radio family.

A remarkable lack of radio responses characterized a series of flares which occurred on 1957 September 16, 17 and 18. Because most of these flares were

SID-associated, more than half of them (about 10) could have been expected to give rise to a radio response. However, only two of them did so. A series of rather insignificant flares which occurred on 1956 May 18 was remarkable for the relatively important associated radio events. An additional peculiarity was the very regular and characteristic shape of the intensity profile of the associated bursts at frequencies near 500 Mc/s The latter circumstance is a further indication that special conditions have governed the radio burst production.

A further application of the notion radio importance lies in the identification of flares that are to be held responsible for sudden commencement geomagnetic storms. It has already been shown by Dodson & Hedeman* that the association of a flare with a major burst at metre wavelengths has a bearing on the probability of that flare being followed by a geomagnetic disturbance. Thanks to the fact that towards the beginning of the I.G.Y. the observational material has become practically complete, we now can state more generally, that there exists a rather close connection between the occurrence of flares of great radio importance and the sudden commencements of major geomagnetic storms.

A major fraction (about 60 per cent) of the major s.c. storms can be related to flares of radio importance 3 or 3+ which preceded by 20–55 hours. Conversely, about the same percentage of the radio importance 3+ flares are followed by a s.c. geomagnetic storm after 1 or 2 days.

Radio importance 3+ flares occur at least three times as frequent as optical importance 3+ flares. The latter flares, which generally also have radio importance 3+, are known to be almost invariably followed by a geomagnetic storm. Most storms, however, are to be ascribed to less important optical flares. Preliminary evidence seems to favour the radio importance above the optical importance as a criterion for the prediction of geomagnetic storms. Those sudden commencements which were not followed by an appreciable geomagnetic disturbance generally could not be related to a particular flare of either great radio or optical importance.

In conclusion we may state, that the notion of radio importance is at least as useful as the optical importance. With regard to solar–terrestrial relationships a flare certainly can more adequately be characterized by both optical and radio importance than by optical importance alone.

Discussion

C. Warwick: We have found that the occurrence of fade-out (SID) is strongly associated with the occurrence of a flare-associated burst at high frequency (decimetre wavelengths), and only weakly related to bursts at metre wavelengths. The occurrence of geomagnetic disturbance, on the other hand, relates more strongly to bursts at metre wavelengths.

We also find a significant relation between the optical importance of the flare and the occurrence in the radio burst of "second part" or type IV.

T. Hatanaka: We have also found that the long-wave outbursts are more effective in the production of magnetic storms.

H. Zirin: The results mentioned by Drs Warwick and Hatanaka fit in with the following facts: certain flares appear as tight bright loops low in the corona with no expulsion of material, but very high excitation. These ought to produce radio

* H. W. Dodson and E. R. Hedeman. *J. Geophys. Res.* 63, 77 (1958).

noise only at high frequencies, and would produce the U.V. or Lα to cause an SID. The other type involve expulsion of a surge or spray, which would produce emission at low frequencies high in the corona and subsequently (in unknown fashion) a geomagnetic storm.

Nouveaux résultats sur la période solaire de 400 ans

F. Link

Poursuivant nos recherches sur le climat et sur l'activité solaire dans le passé historique nous avons pu confirmer et approfondir nos résultats antérieurs (Link & Linkovà, 1955; Link, 1955, 1955a, 1956) par le materiel d'observation enrichi de plusieurs sources et traité d'une façon plus détaillée.

Dans la partie climatique basée sur les découvertes des comètes une nouvelle réduction plus détaillée a confirmé la périodicité de 400 ans dans la nébulosité nocturne en montrant en plus le dédoublement des maxima qui se rencontre sur tous les 5 maxima obtenus depuis le début de notre ère. Nous ne parlerons pas ici de la confirmations de ces résultats par les recherches historiques de Brooks (1950, 1954) mais nous voudrions attirer l'attention sur une extension de notre méthode cométaire aux autres phénomènes astronomiques tels que les éclipses de Soleil et de Lune et les météores. Entre les années 900 et 1 100 de notre ère il a eu lieu une amélioriation considérable de la nébulosité nocturne très bien visible sur notre courbe cométaire. Par un concours favorable des circonstances nous avons pu réunir dans le même intervalle de temps une série assez homogène des observations européennes et chinoises des éclipses de Soleil et de Lune (Riccioli 1651, Gaubil 1732, Struyck 1740, Cinzel 1882, 1883, 1884, 1886, 1899, 1918) de même qu'une série des observations chinoises des étoiles filantes (Biot 1948) qui montrent vers l'an 1 000 le même phénomène que les comètes (Figure 1) confirmant ainsi la réalité de nos résultats cométaires.

Nos recherches de l'activité solaire basées sur les anciennes observations des aurores polaires principalement d'origine occidentale ont pu être enrichies de 30 per cent environ par les observations orientales contenues surtout dans les deux listes de Matsushita (1956). L'ensemble de toutes ces observations montre de nouveau la périodicité de 400 ans des aurores avec les minima situés aux minima de nébulosité ce que permet d'affirmer que ces variations ne sont pas d'origine climatique mais solaire. Ainsi l'activité solaire possède une période séculaire de 400 ans qui influe sur le climat et en particulier sur la pluie (Brooks) et sur la nébulosité nocturne (Link).

Les courbes moyennes de la nébulosité et des aurores (= l'activité solaire) obtenues par la superposition de 5 périodes depuis le début de notre ère sont données sur la Figure 2. Les minima principaux des deux courbes coïncident pratiquement tandis que le reste de la courbe aurorale est fortement déformé par l'influence de la nébulosité. Ainsi les deux maxima auroraux l'un peu après le minimum de la nébulosité et l'autre peu après le maximum ne sont pas forcément situës aux maxima de l'activité solaire étant plutôt l'effet combiné du nombre croissant des aurores et du nombre décroissant des nuits favorables aux observations. Au contraire cette déformation n'est pas à craindre aux minima des deux courbes. Remarquons encore que différentes recherches antérieures dans les domaines solaire (Brooks 1928, Oppenheim 1928) climatique (Craig 1954) et

géomagnétique (Schmidt 1932) ont déjà fait entrevoir la possibilité d'une longue période analogue à la notre sans pouvoir toutefois démontrer cette variation dans l'intervalle de temps suffisant ou d'une manière convainquante.

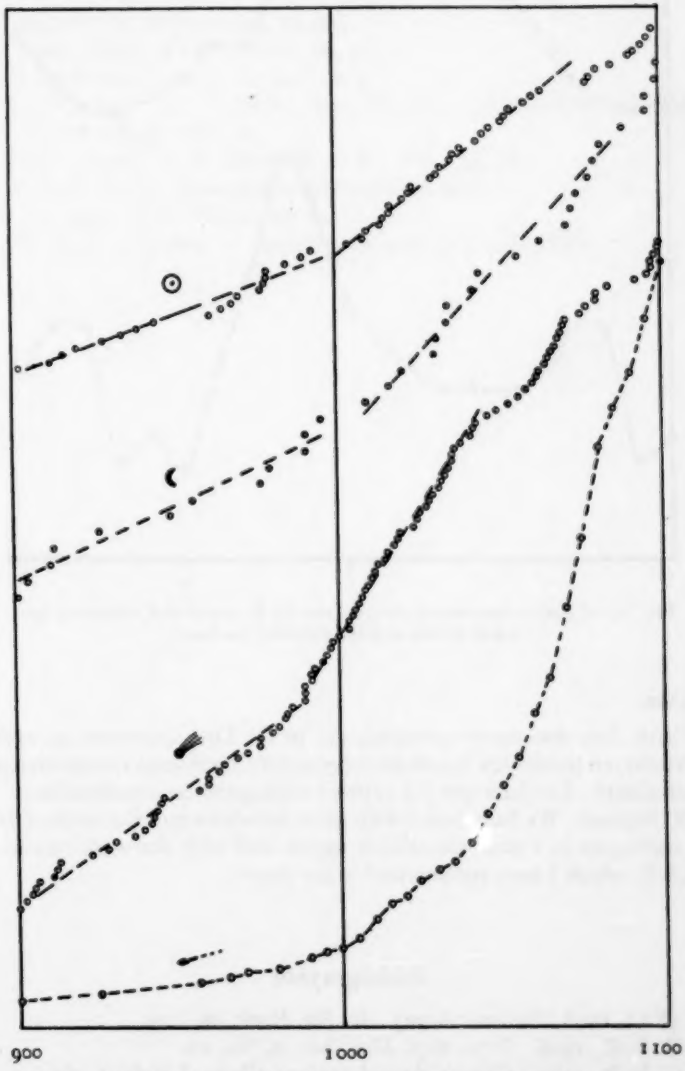

FIG. 1.—Variation du nombre d'observations (cumulatives) des éclipses de Soleil, de Lune, des cometes et des étoiles filantes entre les années 900 et 1100 de notre ère.

D'après tout ce qui precède la période de 400 ans trouvée par les nouvelles méthodes introduites dans la climatologie historique doit figurer définitivement dans l'état actuel de nos connaissances sur l'activité solaire en se rangeant à coté des periodes déjà connues de 11 et de 80 ans.

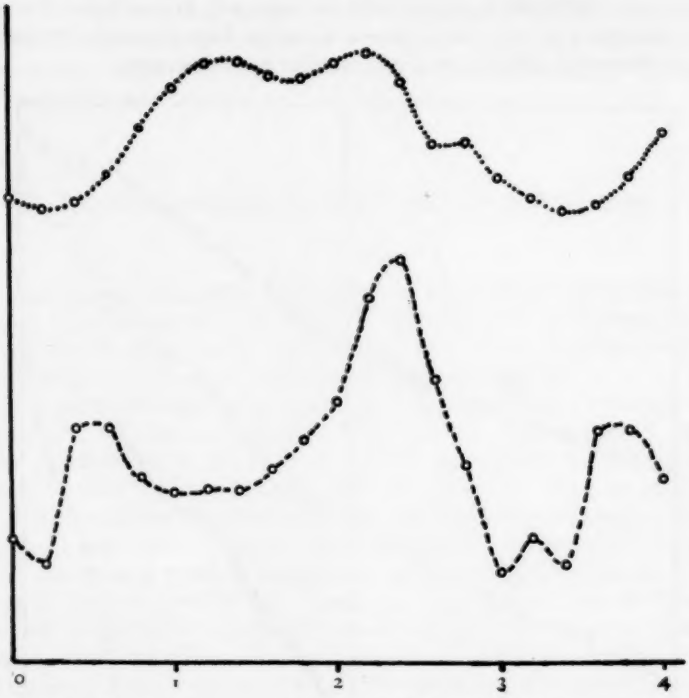

FIG. 2.—Courbes moyennes de 400 ans de la nébulosité nocturne (en haut) et des aurores polaires (en bas).

Discussion

E. Vassy: Les documents présentés par le Dr Link apportent un appui que j'ai exprimées, en particulier la courbe representant l'influence des aurores polaires sur la nebulosité. Les faits que j'ai exposés expliquent ses constatations.

M. S. Eigenson: We have heard with great satisfaction of the work of Dr Link and his colleagues in Czechoslovakia: it agrees well with the work carried out in the U.S.S.R. which I have summarized in my paper.

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Book Reviews

External Gravity Field of the Earth and Fundamental Constants Involved

I. Zhongolovitch

(Published by the U.S.S.R. Academy of Science, Institute of Theoretical Astronomy, Vol. III, Moscow 1952. pp. 125. In Russian)

In 1952 this was an important paper on the determination of the form of the geoid from observed values of gravity, but copies do not seem to have reached many Western geodesists. Seven years later it still contains much that is of value.

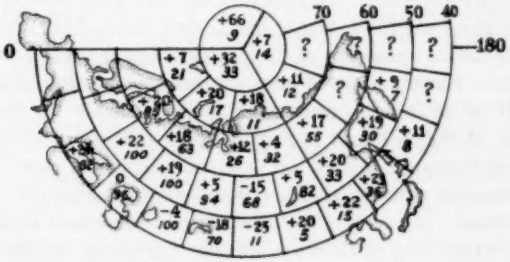
The paper starts with a list of gravity values at 62 National Base stations and other common points (mostly old comparisons: nothing since 1941 except a few Woollard 1948), which in 1959 is out of date. The author then devises a scheme of 412 equal-area "ten-degree squares" covering the whole Earth. Starting with 36 such squares between Lat 0° and approximately 10° N, the numbers in each 10° belt are successively reduced to 34, 32 . . . 15 and 9 (70°–80°) with three triangular areas at the Pole, and exact equality of area is secured by slight departures of the 10-degree latitude limits from exact multiples of 10° (pages 25–28).

Pages 34–41 then give mean gravity anomalies where available for each of these squares (150 squares in North latitudes and 54 in South), with a column (n) showing the percentage of single-degree squares containing data within each 10-degree square. Two anomalies are given for each 10°-square; namely Δg_1 which is the simple mean of the mean $(g-\gamma_A)$ in each of the n 1°-squares: and Δg_2 which is the value given for the mean height of the area from a straight line fitted to the 1° square means when plotted against average station heights. The procedure is described in detail on pp. 28–33 of which an English translation exists. On pp. 34–41 the table gives $\bar{h}$ and $\bar{h}^1$ the average height of land and/or depth of sea in km in each 10-degree square: α and β the percentage areas covered by land and sea: and b and b^1 the coefficients in the expression $g-\gamma_A = a+b\bar{h}$ for land and sea respectively. This procedure is similar to that followed by H. Jeffreys (1941), and is a very effective method of meaning $g-\gamma_A$ from uneven data over a small area such as a degree-square. Over a 10-degree square it is necessarily less effective, as the variations in b show, but it is thought to be the best that can easily be done without computing orographic, isostatic or other such corrections. The Figure gives values of Δg_2 (and n) over parts of Asia and the Arctic where information has hitherto been scanty. For those who have not had access to Russian data, this must be the most valuable part of the book.

The author then makes a spherical harmonic analysis for Δg_1 and Δg_2 for all harmonics up to degree 8, and gives the 81 constants for each solution in Table 8 on pp. 56–57. The "impossible" first- and second-degree harmonics are not suppressed, but come out reasonably small (0·1, 0·5, 1·6, 0·1 and 0·6 mgal for Δg_2). Pages 58 and 59 sum the terms of the analysis for each 10-degree square (for comparison with the original data) and Figures 5 and 6 show the Δg_2 result graphically.

An analysis based on data available in 1952 must clearly be in some respects out of date in 1959, but the order of magnitude of the different terms is of interest. The volume of observations available, even now, is such that analysis can hardly be expected to give more than a correct order of magnitude after the 3rd degree and P_4.

Pages 67 and 75 give a summary of various deduced "constants" of the Figure of the Earth: equatorial gravity, flattening (1/296·6 from Δg_2), inequality of equatorial semi-axes (199 m with maximum in Long 6° West), difference of moments of inertia, etc.



Anomaly Δg_2 +66
Percentage cover 9

Pages 80–81 give values of the 76 possible harmonic constants for the anomalies in the form of the geoid, namely Table $8 \times (6378/980) \div (n-1)$ metres. The resulting geoidal rise (from Δg_2) is given on page 83 and shown graphically in Figures 7 and 8 at end, Figure 8 being with reference to a tri-axial standard. The factor $(n-1)$ naturally gives emphasis to the lowest harmonics in the gravity analysis, such as the 2nd degree sectorial, and the consequent smoothness of the contours in Figure 7 gives a possibly deceptive appearance of reality to the tri-axial form. The true form of the geoid (when it is known) may or may not show a second-degree variation of ± 150 m above a best-fitting oblate spheroid, but it will also almost certainly show equally significant large random irregularities of much shorter pitch.

The book finishes with a long table (pp. 92–119) giving the average values of each of the 80 harmonic terms p_{10} to $p_{88} \times (\cos \lambda$ or $\sin \lambda)$ in each of the 206 "10-degree" squares of a hemisphere. Exact definitions of p_{nm} and P_{nm} are given on pp. 92 and 120–122. The arrangement of the table, which is not very apparent at first sight, is that for each term the 206 values are serially arranged in 21 rows of 10 columns with 4 blank spaces at the end. The completion of this table and the average heights given earlier for each (equal) "10-degree" square suggest that this scheme of sub-division might usefully be generally adopted by other workers in the same field.

G. Bomford.

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Celestial Mechanics

E. Finlay-Freundlich

(150 pp., Pergamon Press, 1958)

A new book on celestial mechanics is timely in view of the revival of interest in the subject occasioned by the launching of artificial satellites and space probes. Outstanding though the results are that have already been obtained from analyses of the motion of close satellites, many fascinating theoretical problems remain to be studied, while the examination of the orbits of space probes will be very intricate. Furthermore, the methods of analysis hitherto developed for the planetary and lunar problems are not necessarily the most suitable for the study of close satellites because of the different approximations that are appropriate. In these circumstances one looks for a book on celestial mechanics that will give a broad fundamental account of principles rather than a very detailed description of planetary or lunar methods. However, the stimulus for Professor Finlay-Freundlich's book was not the launching of satellites but his interest in double stars and so the reader will find a number of omissions, particularly of Lagrange's planetary equations, which rather reduce its value for satellite studies. But to offset this, there is a valuable account of the Hamilton–Jacobi theory and of the use of canonical variables in dealing with small perturbations. Anyone concerned with celestial mechanics will profit from a study of these chapters.

Professor Finlay-Freundlich applies the method of canonical co-ordinates to the analysis of the motion of a pair of deformable stars and shows how the dynamical results support the idea that stars are polytropes with n equal to 3. Geodesists will be interested in the treatment of the Clairaut theory.

There are also chapters on the n-body problem (this contains some general results relevant to the orbits of space probes) on the general features of the motion of the Moon, and on effects in general relativity.

To summarize, this is not a manual of methods that can be directly applied to planetary, lunar or close satellite problems, but the discussion of basic ideas and the general approach to the problems will be found illuminating and helpful.

A.H.C.

РАСПРОСТРАНЕНИЕ ГЕОМАГНИТНЫХ ВОЗМУЩЕНИЙ ЧЕРЕЗ АТМОСФЕРУ И МЕЖПЛАНЕТНОЕ ПРОСТРАНСТВО

Дж. Х. Пиддингтон

Развита теория распространения медленно меняющихся электромагн. возмущений через частично ионизированную газовую среду и применена к земной атмосфере и межпланетному пространству. Среда принимается состоящей из двух сосуществующих отдельных газов—электронно-ионная плазма и нейтральные атомы—движущихся до некоторой степени независимо. Численные результаты даны для модели атмосферы, простирающейся до расстояния в несколько земных радиусов.

1. До расстояния в несколько сот километров среда имеет свойства, для колебаний всех периодов от 1 до 10^4 сек., неподвижного проводника и диспергирующей среды.

2. Выше 10^3 км возмущения распространяются в виде гидромагнитных волн в чисто ионной плазме. Потери (энергии) в этой области малы, но распространение вероятно осложняется процессами преломления и частичного отражения.

3. Токи, ответственные за все наблюдаемые геомагнитные возмущения, должны течь в слоях, расположенных ниже 1000 км.

4. Обсуждена проблема проникновения солнечного газа в земное поле.

5. Обсуждены некоторые прежние теории главной фазы магнитных бурь с внезапными началами (SC) и сделано альтернативное предположение. Таким образом часть силовых линий земного поля перемещается вдоль линии Солнце-Земля.

6. Некоторые свойства микропульсаций объяснены при помощи теории распространения гидромагнитных волн.

7. «Эффективная» проводимость (σ_3) атмосферы на расстоянии до нескольких земных радиусов имеет максимум на высоте 100 км, затем быстро убывает до величины 10^{-16} эл.-магн. ед. или меньше, и более не возрастает.

ВРЕМЕНА ПРОБЕГА ВОЛНЫ PKP ДО 145°

Б. А. Болт

Проанализировано (по данным Международной сейсмологической сводки) около 900 времен вступления волны PKP в интервале расстояний $110° < \Delta < 145°$ при 24 южных землетрясениях.

Получены новые данные в пользу того, что времена, указанные в таблицах Джеффриса-Буллена для ветви ***DE*** на расстояниях до 140° следует увеличить примерно на секунду. Нет указаний на заметную разницу времен пробега PKP для Европы и Северной Америки в этом интервале.

На расстояниях $135° < \Delta < 145°$ наблюдения над PKP группируются на временах, меньших чем для ветви ***DE***, согласно формуле: $t = 19$ мин 29,1 сек $+ 2{,}69\ (\Delta - 141°{,}6)$ сек. Быть может это соответствует ветви *BC* волны PKP и ее переходу в диффрагированную волну, связанную с соответствующей точке *B*.

ВНЕШНЕЕ ГРАВИТАЦИОННОЕ ПОЛЕ ВРАЩАЮЩЕГОСЯ СФЕРОИДА ДО ЧЛЕНОВ ПОРЯДКА e^3

А. Х. Кук

Теория внешнего гравитационного поля вращающегося тела, ограниченного сфероидальной эквипотенциальной поверхностью, содержит члены порядка e^3. Потенциал выражается функциями, которые ортогональны внешней поверхности сплющенного сфероида. Найдено, что члены порядка e^3 в гравитационной формуле и в соотношениях между J, e и m обычно очень малы.

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